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THE VELOCITY OF SOUND IN ROCKS AND GLASSES AS A FUNCTION OF TEMPERATURE[1]

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ABSTRACT

Measurements show that the velocity of sound in Pyrex and in silica glass rises with increasing temperature and that these substances support vibrations substantially without loss to temperatures above 700° C. Measurements on rock samples show a decrease in velocity with increasing temperature, only part of which is reversible as the material cools. This permanent reduction in velocity is explained by differential thermal expansion. The reversible temperature effect on velocity of sound is less than 1 per cent per hundred degrees in granite, and about 5 per cent per hundred degrees in diabase. The geophysical significance of these results is discussed.

INTRODUCTION

Reasoning about the composition of the earth's crust requires a knowledge of the way in which the elastic properties of rocks vary under combined high temperature and high pressure. From this information curves of wave-velocity with depth in the earth can be computed, assuming that the crustal layers are made up in various ways. By comparing such curves with the known variation of velocity with depth, obtained from seismic data, identification of the layers should be possible. Laboratory investigation has resulted in the measurement of the compressibility, rigidity, Poisson's ratio, and Young's

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modulus of rock samples at atmospheric pressure.[2] The compressibility has been measured as a function of hydrostatic pressure,[3] and recently the compressibility of glasses has been measured under combined high temperature and high pressure.[4] The effect of pressure on one modulus is thus known, while the effect of temperature operating alone has never been subjected to investigation. This paper reports measurements of the velocity of sound in rocks and glasses as a function of temperature, in the hope that some light may be cast on geophysical and geological problems.

METHOD OF MEASUREMENT

A brief review of methods will show that those commonly employed to measure elastic temperature coefficients in metals and in glasses are not suitable for work with rock samples. Static methods[5]

[2] W. A. Zisman, "Young's Modulus and Poisson's Ratio with Reference to Geophysical Applications," *Proc. Natl. Acad. Sci.*, Vol. XIX (1933), pp. 653–65; "Compressibility and Anisotropy of Rocks at and near the Earth's Surface," *ibid.*, pp. 666–79; "Comparison of the Statically and Seismologically Determined Elastic Constants of Rocks," *ibid.*, pp. 680–86; J. M. Ide, "Comparison of Statically and Dynamically Determined Young's Modulus of Rocks," *ibid.*, Vol. XXII (1936), pp. 81–91; "The Elastic Properties of Rocks: A Correlation of Theory and Experiment," *ibid.*, pp. 482–96.

[3] L. H. Adams and R. E. Gibson, "The Cubic Compressibility of Certain Substances," *Jour. Wash. Acad. Sci.*, Vol. XXI (1931), pp. 381–90; P. W. Bridgman, "The Thermal Conductivity and Compressibility of Several Rocks under High Pressures," *Amer. Jour. Sci.*, Vol. VII (1924), pp. 81–102; "The Compressibility of Several Artificial and Natural Glasses," *ibid.*, Vol. X (1925), pp. 359–67; L. H. Adams and E. D. Williamson, "On the Compressibility of Minerals and Rocks at High Temperatures," *Jour. Franklin Inst.*, Vol. CXCV (1923), pp. 475–530.

[4] Francis Birch and R. R. Law, "Measurement of Compressibility at High Pressures and High Temperatures," *Geol. Soc. Amer. Bull. 46* (1935), pp. 1219–49; Birch and R. B. Dow, "Compressibility of Rocks and Glasses at High Temperatures: Seismological Application," *Geol. Soc. Amer. Bull. 47* (1936), pp. 1235–55.

[5] H. L. Dodge, "The Change in the Elasticity of a Copper Wire with Current and External Heating," *Phys. Rev.*, Vol. II (1913), pp. 431–49; "The Change in the Elasticity of a Mild Steel Wire with Current and External Heating," *ibid.*, Vol. V (1915), p. 373; "The Change in the Elasticity of Aluminum Wire with Current and External Heating," *ibid.*, Vol. VI (1915), pp. 312–18; "Young's Modulus of Drawn Tungsten and Its Variation with Change of Temperature," *ibid.*, Vol. XI (1918), pp. 311–15; W. Mobius, "Der Torsionsmodul des Nickels bei höheren Temperaturen unter gleichzeitiger Magnetisierung," *Phys. Zeitschr.*, Vol. XXXIII (1932), p. 417; C. H. Lees, J. P. Andrews, and L. S. Shave, "The Variation of Young's Modulus at High Temperatures," *Proc. Phys. Soc. London*, Vol. XXXVI (1924), pp. 405–16; J. P. Andrews, "The Variation of Young's Modulus at High Temperatures," *ibid.*, Vol. XXXVII (1925), pp. 169–77.

of all kinds are unreliable for the measurement of imperfect elastic materials such as rocks. Piezoelectric driving devices[6] require much power and involve the complications inherent in coupled circuits. It is difficult to cut rocks thin enough to obtain transverse vibrations, although metals have been studied in this way.[7] A rock sample may be driven magnetostrictively by wrapping nickel foil around it and placing it in a coil carrying alternating current of the proper frequency, but the fading of magnetostriction with rising temperature eliminates this method for the present study. These difficulties may be avoided by the use of the electrostatic method, recently described by the author as a general means of measuring the velocity of sound in substances which are available in cylindrical samples.

The method, described in detail in the *Review of Scientific Instruments* (October, 1935), may be outlined briefly as follows: The rock specimen, with metal foil cemented to its lower end, is placed upright on a thick steel disk. The foil and steel surfaces are separated by a thin sheet of mica. An alternating voltage of variable frequency is applied across the electrical condenser thus formed. When the frequency of the voltage resonates with the natural longitudinal frequency of the specimen, the latter vibrates with increased amplitude and radiates sound. This is detected by ear or by a piezoelectric crystal receiver. When the cylinder vibrates in its fundamental mode, its center is a node and its ends are positions of maximum amplitude. The sample is a half-wave-length vibrator, and the velocity of sound in it is equal to the resonant frequency multiplied by twice the length of the cylinder. The method is simple and rapid, and accurate within a few tenths of 1 per cent.

This method is readily adapted to the measurement of the velocity of sound at elevated temperatures. The specimen electrode of tin or lead foil is replaced by silver foil, to which a silver connecting wire is brazed. The specimen is placed in a cylindrical electric furnace of conventional design. The cover of this is removed when the sample is set into vibration, and the frequency of maximum acoustic re-

[6] J. Zacharias, "The Temperature Dependence of Young's Modulus for Nickel," *Phys. Rev.*, Vol. XLIV (1933), p. 116; L. Balamuth, *ibid.*, Vol. XLV (1934), p. 715.

[7] K. R. Koch and R. Dieterle, "Die Elastizität Einiger Metalle und Legierungen bis zu Temperaturen, die ihrem Schmelzpunkt naheliegen," *Ann. der Physik*, Vol. LXVIII (1922), p. 441.

sponse is determined by ear or by means of a piezoelectric microphone and amplifier placed a foot or two above the top of the furnace. Most of the resonant frequencies fall well within the audible range and become lower as the temperature rises. A few resonances are above audibility, detected by means of a heterodyne sound-receiver, loaned for this research by Professor G. W. Pierce.

The alternating voltage required to set the specimens into vibration was supplied by a General Radio Type 377B low-frequency oscillator, with one stage of power amplification, which raised the output to 150 volts. A steady polarizing voltage of about 600 volts was provided by a simple vacuum-tube rectifier. The current requirement is negligible, since the impedance of the specimen condenser is of the order of 1,000,000 ohms. With maximum applied voltage, the amplitude of the stress applied to the specimen end is about 18 grams per square centimeter.

Two furnaces were employed. The interior of the larger one was cylindrical in shape, 3 inches in diameter and 16 inches in height. This was constructed to hold specimens 2 inches in diameter and 10 inches long. The smaller furnace had an interior $1\frac{1}{2}$ inches in diameter and 12 inches in height. The desired temperatures were held constant by means of a bridge-type thermostat and were measured by means of a chromel-alumel thermocouple fastened to the center of the specimen. In the lower temperature ranges the thermocouple readings were frequently checked with a mercury thermometer. The furnaces and thermostat were constructed under the direction of Dr. Francis Birch.

The method of this research could probably be extended to temperatures above 750° C., at least for some materials. The furnaces were not constructed for higher temperatures, since this seemed adequate for a first survey of the subject.

The rock samples studied were chosen from the group of specimens prepared for the Harvard Committee on Geophysical Research. The elastic properties of these samples have been extensively investigated by the author and by W. A. Zisman.[8] The earlier articles contain descriptions of the various rocks, based on petrographic data.

[8] Zisman, *op. cit.*; Ide, *op. cit.*

Four of the specimens—Vinal Haven diabase No. 1, Quincy surface granite No. 8, French Creek norite No. 3, and quartzitic sandstone No. 2 from Pennsylvania—were in the form of cylinders 2 inches in diameter and 10 inches long. The marble sample from Danby, Vermont, was square in cross section, 2 inches on a side, and 12.8 inches long. The sample of mild carbon steel was a round rod 1 inch in diameter and 13.3 inches long. These were heated in the large furnace.

Nine specimens were heated in the small furnace. They were as follows: Sudbury norite No. 6 and No. 4, $1\frac{1}{8}$ inches in diameter and 8 inches long; Solenhofen limestone, $\frac{5}{8}$ inches in diameter and 6.68 inches long; Lipari obsidian, $\frac{5}{8}$ inches in diameter and 4.04 inches long. The foregoing samples have all been described in earlier papers. To them were added the following glass specimens: (1) A soft glass, obtained from Professor P. W. Bridgman and described in his papers as "Glass C." This glass was made at the Corning Glass Works and is best described as a soda-potash-lime silicate. The silica content is 70 per cent, and the density 2.507 gm/cm^3. (2) A sample of "Pyrex chemical resistance glass," also from the Corning Glass Works. This is a soda-alumina boro-silicate, containing 81 per cent of silica. Its density is 2.244 gm/cm^3. (3) A rod of fused silica glass, of density 2.207 gm/cm^3, was obtained from the Thermal Syndicate. The Pyrex sample was $\frac{3}{4}$ inch, and the others $\frac{1}{2}$ inch, in diameter. Two metal rods—one of copper, 1 inch in diameter; and one of elinvar (37 per cent Ni, 8 per cent Cr, balance Fe), $\frac{1}{2}$ inch in diameter—were included among the samples. The three glass and two metal specimens were all 8 inches in length.

The quantity measured in these experiments requires further definition to avoid possible confusion. The velocity of sound, as ordinarily defined for solids and listed in the handbooks, is equal to $\sqrt{E/\rho}$, where E is Young's modulus of elasticity, and ρ the density of the material. Strictly, this is the velocity of a longitudinal wave propagated in an infinitely long rod. In practice, if the length of the rod is five or more times its diameter, as in the present experiments, the measured velocity differs from that in an infinite rod by less than the experimental errors. From the velocity measured in this way Young's modulus of elasticity may be computed, and is equal to the

square of the velocity multiplied by the density. Since the density does not vary more than a few per cent over the range of these experiments, the velocity changes recorded are primarily due to the effect of temperature on Young's modulus of elasticity.

The recorded values of velocity are not corrected for the effect of thermal expansion during the runs. The length of the sample is assumed constant in computing velocity from resonance frequency. The length actually increases with temperature, by amounts varying from 0.04 per cent in silica, 0.25 per cent in Pyrex, to perhaps 1 or 2 per cent in rocks like quartzitic sandstone and granite. In the glasses this effect may be neglected, since it is below the limit of experimental error. In the rocks, the velocity changes with temperature are so large that 1 or 2 per cent correction for thermal expansion would not be visible on the graphs.

In geophysical applications, the term "velocity of a longitudinal wave" refers to the velocity of sound in a medium indefinitely extended in all directions. This is larger than the velocity in a long rod by a factor which is numerically between unity and seven-fifths for all known rocks. The exact evaluation of this factor requires a knowledge of one other elastic constant in addition to Young's modulus. The reader is referred to treatises on elasticity for details.[9]

RESULTS FOR STEEL, COPPER, AND ELINVAR

The results of the measurements will be described in the order of increasing complexity. The samples of steel, copper, and elinvar were studied with the object of demonstrating the method on known materials.

The curves for the metals are given in Figure 1, and the corresponding data in Table 1. The velocity of sound in steel shows a systematic decrease as the temperature rises, and the rate of decrease becomes gradually greater. The curve of velocity against temperature is thus slightly concave downward. The velocity decreases 15.7 per cent as the temperature rises to 700° C. As may be seen by consulting the papers of Koch and Dieterle, of Lees, Shave, and Andrews, or of H. L. Dodge, this type of curve is characteristic of

[9] A. E. H. Love, *The Mathematical Theory of Elasticity* (4th ed., Cambridge, 1934).

most materials. If we compute from our data the temperature coefficient of the velocity of sound in steel, for temperatures below

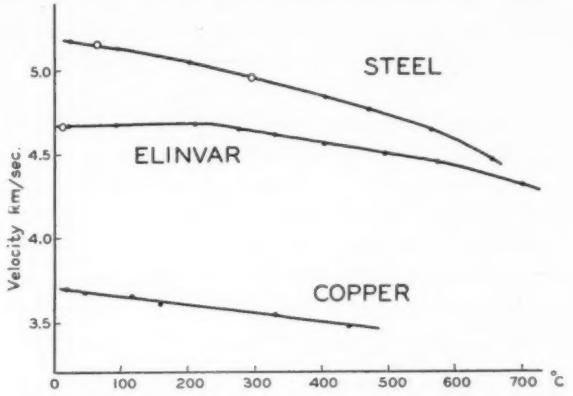

FIG. 1.—The velocity of sound in steel, copper, and elinvar as a function of temperature.

TABLE 1*

THE VELOCITY OF SOUND AS A FUNCTION OF TEMPERATURE IN VARIOUS MATERIALS

STEEL		COPPER		ELINVAR		LIMESTONE		PYREX		SILICA		OBSIDIAN		SOFT GLASS	
T*	v*	T	v	T	v	T	v	T	v	T	v	T	v	T	v
25	5.170	47	3.670	23	4.670	20	4.920	19	5.305	18	5.730	20	5.433	55	5.288
107	5.120	119	3.650	93	4.675	104	4.886	93	5.365	154	5.800	97	5.443	120	5.262
210	5.036	20	3.700	214	4.685	213	4.756	222	5.410	282	5.843	228	5.447	226	5.200
299	4.940	161	3.600	278	4.645	179	4.820	313	5.450	437	5.894	296	5.454		
408	4.837	332	3.540	333	4.603	108	4.860	546	5.494	778	5.973	20	5.433		
472	4.765	445	3.480	405	4.560	276	4.660	495	5.504	698	5.970				
411	4.833			498	4.500	342	4.610	400	5.500	556	5.950				
568	4.640			576	4.450	448	4.545	50	5.365	32	5.734				
660	4.467			702	4.310	400	4.580	168	5.445						
296	4.040			98	4.690	305	4.640	325	5.470						
65	5.155			14	4.670	21	4.793	571	5.485						
						101	4.726	595	5.470						
						198	4.660	621	5.447						
						302	4.630	638	5.400						
						385	4.580	649	5.390						
								660	5.370						
								670	5.330						
								684	5.290						
								701	5.210						
								567	5.506						
								20	5.380						

* T, temperature in degrees centigrade; v, velocity of sound in kilometers per second.

100° C., we obtain −1.34 per cent per hundred degrees. This compares closely with the value −1.37 per cent per hundred degrees

published by Professor G. W. Pierce from measurements with magnetostriction oscillators.[10]

The curve for copper requires no particular comment. Its shape is similar to that for steel, and its slope yields a temperature coefficient of velocity of sound equal to -1.38 per cent per hunderd degrees.

The curve for elinvar shows strikingly the peculiar property of this material, viz., a very small temperature coefficient of elasticity at ordinary temperatures. From room temperature to 230° C. this coefficient averages 0.134 per cent per hundred degrees and is positive in sign. Above that temperature the curve breaks abruptly and has the negative slope and downward curvature usually associated with metals. Such abrupt changes in elasticity at the Curie point are discussed theoretically in a recent paper by M. Kersten.[11]

VELOCITY OF SOUND VERSUS TEMPERATURE IN GLASSES

We have measured the velocity of sound as a function of temperature in four markedly different types of glass: soda-lime soft glass, Pyrex chemical resistant glass, fused silica glass, and Lipari obsidian. The data are given in Table 1, and the corresponding curves in Figure 2. Several investigators have measured by static methods the elasticity of soft glasses as a function of temperature,[12] and some data have been published for silica glass. So far as the author is aware, the data presented here for Pyrex and obsidian are the first of this type available.

We now wish to consider the results in detail. The velocity of sound in the sample of soft glass decreases steadily from 5.3 km/sec as the temperature rises to 365° C. The velocity curve is concave downward, with an average slope of -1.31 per cent per hundred degrees. Above 330° C. the decrement of the vibrations begins to increase rapidly, and at 365° C. resonance is spread over a broad

[10] G. W. Pierce, "Magnetostriction Oscillators," *Proc. I. R. E.*, Vol. XVII (1929), p. 42.

[11] M. Kersten, "Über den Temperaturkoeffizienten des Elastizitätsmoduls ferromagnetischer Stoffe," *Zeitsch. f. Phys.*, Vol. LXXXV (1933), pp. 708–17.

[12] A. K. Gerhard Schulze, "Über die Temperaturabhängigkeit des Elastizitätsmodulus einiger Glassorten," *Zeitsch. f. Phys.*, Vol. LXIX (1931), pp. 456–58. See also n. 5.

band of frequencies. At yet higher temperatures the material becomes unable to support resonant vibrations, owing to excessive internal friction losses.

The sample of Pyrex chemical resistance glass shows an increase in the velocity of sound with temperature. The velocity rises smoothly from 5.3 km/sec at room temperature to 5.49 km/sec at 550° C., decreases to 5.46 km/sec at 600° C., and then more rapidly to 5.2 km/sec at a temperature of 700° C. The average temperature coefficient of velocity below 400° C. is +0.8 per cent per hundred

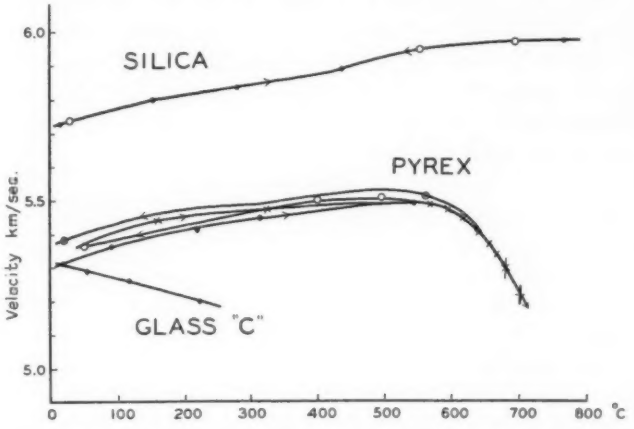

FIG. 2.—The velocity of sound as a function of temperature in silica, Pyrex, and soft glass.

degrees. The decrement of the vibrations shows no noticeable variation until about 650° C., and then increases rapidly until the system is practically nonoscillatory above 700° C. At still higher temperatures no resonance frequency can be identified. On cooling the material to room temperature, the velocity of sound returns to a value 1.3 per cent higher than at the beginning of the measurements. Irreversible effects of this magnitude may also be observed in metals, where they are attributed to the removal of internal strains by annealing.

The sample of silica glass, or fused quartz, also shows an increase in the velocity of sound with rising temperature. The velocity at

room temperature is 5.72 km/sec, and this increases smoothly to 5.96 km/sec at 780° C. The curve of velocity against temperature is very slightly concave downward and has an average slope of +0.55 per cent per hundred degrees. The temperature coefficient of velocity is thus positive and about two-fifths as large as that of steel. On cooling the material to room temperature, the velocity returns to its original value. The silica rod shows no diminution in ability to sustain vibrations at 780° C., and there is no decrease in the intensity of the emitted sound or in the sharpness of resonance. The furnace did not permit observations on this sample at higher temperatures.

The Lipari obsidian was the only sample of natural rock glass available for these experiments. Because of the shortness of the specimen, its natural frequency was very high (26,500 cycles per second), and the measurements proved too difficult to carry above 300° C. The velocity of sound is 5.435 km/sec at 20° C., and this increases one-third of 1 per cent as the temperature rises to 300° C. This gives a positive temperature coefficient of velocity equal to +0.11 per cent per hundred degrees. The curve for obsidian is omitted from Figure 2, since it would be practically a horizontal line.

An interesting result of these measurements is the abnormal *increase* in the velocity of sound with temperature observed in Pyrex chemical resistance glass, silica glass, and natural obsidian glass. This effect was observed in silica by C. H. Lees, J. P. Andrews, and L. S. Shave.[13] Their results are rather discordant, because of the limitations of the static method; but several runs on silica show an increase in elasticity. Recent measurements by Birch and Law[14] show an anomalous decrease in the compressibility of certain glasses with rising temperature. The effect is most marked in silica but appears also at temperatures above 200° C. in obsidian and in diabase glass. Anomalous behavior under high pressure has been observed by P. W. Bridgman, by L. H. Adams and R. E. Gibson in silica glass and other glasses (such as Pyrex) of high silica content. It seems reasonable that the abnormalities which we have found in the velocity of sound as a function of temperature should be correlated with unusual behavior of the compressibility as a function of temperature and of pressure.

[13] *Op. cit.* [14] *Op. cit.*

By combining the compressibility measurements of Birch and Dow[15] with our measurements of sound velocity in a long bar, and assuming isotropy of the materials, it is possible to compute the wave-velocity in an infinite medium of each of the following sub-

TABLE 2

LONGITUDINAL WAVE VELOCITIES COMPUTED FROM MEASURED VALUES OF YOUNG'S MODULUS AND COMPRESSIBILITY*

T	β	E	V
Pyrex Glass			
0	31.05	6.300	5.51
100	30.13	6.475	5.58
200	28.05	6.577	5.69
300	27.77	6.675	5.72
400	28.63	6.730	5.70
Obsidian			
0	27.60	6.940	5.66
100	27.90	6.963	5.66
200	27.83	6.973	5.66
300	27.47	6.989	5.68
Soda-Lime Glass			
0	24.02	7.03	5.67
100	25.01	6.97	5.60
200	26.54	6.83	5.49
300	28.07	6.57	5.38
Solenhofen Limestone			
0	12.85	6.37	6.48
100	14.18	6.19	6.25
200	15.40	5.94	6.03
300	16.52	5.57	5.82
400	17.03	5.47	5.75
300	14.43	5.56	6.10
200	13.20	5.66	6.32
100	12.31	5.81	6.51
0	11.66	6.00	6.67
Silica Glass			
0	25.4	7.22	6.00
100	25.7	7.36	6.03
200	24.8	7.47	6.09
300	24.2	7.54	6.13
400	24.5	7.63	6.15
500	.....	7.75	.....
600	.....	7.83	.....
700	.....	7.86	.....
800	.....	7.87	.....

* T, temperature in degrees centigrade; β, compressibility $\times 10^{13}$ square centimeters per dyne, from measurements by F. Birch and R. B. Dow; E, Young's modulus $\times 10^{-11}$ dynes per square centimeter, computed from velocity measurement by J. M. Ide; V, velocity of a longitudinal elastic wave in an extended medium, in kilometers per second.

stances: soft glass, Pyrex, silica, obsidian, and Solenhofen limestone. This is of interest since the longitudinal velocity in an infinite medium is the elastic constant most easily derivable from seismological data.

The results of these computations are given in Table 2. Here we

[15] *Op. cit.*

list at 100°- temperature intervals: (1) the compressibility, β, taken from the measurements of Birch and Dow; (2) Young's modulus of elasticity, E, computed from our velocity measurements; and (3) the

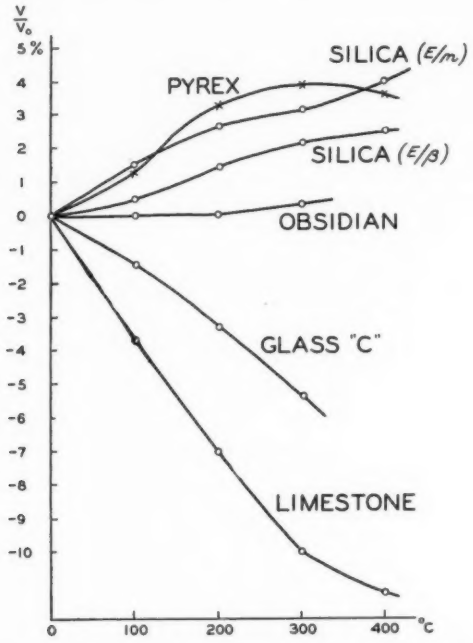

FIG. 3.—The percentage change with temperature of the longitudinal wave velocities in various substances. Velocities are computed from measured Young's modulus and compressibility.

longitudinal wave velocity in an infinite medium, V, computed from the formula

$$V^2 = \frac{3}{\rho\beta}\,\frac{3 + E\beta}{9 - E\beta}.$$

Figure 3 shows curves of wave-velocity as a function of temperature, at atmospheric pressure, using the data of Table 2, plotted as percentage change from the velocity at 0° C. For silica we give two wave-velocity curves, one set of values computed from E and β,

using the formula above, and another set of values computed from E and μ, using the formula

$$V^2 = \frac{\mu}{\rho} \frac{4 - \frac{E}{\mu}}{3 - \frac{E}{\mu}}.$$

Here μ is the modulus of rigidity. The rigidity modulus of silica glass as a function of temperature has been measured by F. Horton,[16] In Table 3, the temperature variation found by Horton has been

TABLE 3

ADDITIONAL ELASTIC PROPERTIES OF SILICA*

T	μ	σ	$E\beta + 3\frac{E}{\mu}$	V
0.....	3.041	0.187	8.95	5.99
100.....	3.071	.198	9.08	6.08
200.....	3.099	.205	9.08	6.15
300.....	3.129	.205	9.05	6.18
400.....	3.153	.210	9.13	6.23
500.....	3.176	.220	.....	6.33
600.....	3.196	.225	.....	6.39
700.....	3.214	.222	.....	6.39
800.....	3.225	0.220	.....	6.38

*T, temperature in degrees centigrade; μ, the modulus of rigidity, after F. Horton and F. Birch, in dynes per square centimeter times 10^{-11}; σ, Poisson's ratio computed from E and μ; V, velocity of longitudinal waves in infinite medium, in kilometers per second, computed from E and μ.

applied to the absolute value of μ at 30°, found by Birch[17] for the same sample of silica glass used for the measurements of the present paper. In a homogeneous, isotropic, and perfectly elastic substance the theory of elasticity requires that

$$E\beta + 3\frac{E}{\mu} = 9.$$

The third column of Table 3 shows how nearly silica glass obeys this formula, as the temperature increases from zero to 400° C. The

[16] F. Horton, "On the Modulus of Torsional Rigidity of Quartz Fibres and Its Temperature Coefficient," *Phil. Trans. Roy. Soc. London*, Ser. A, Vol. CCIV (1905), p. 407.

[17] F. Birch, "Effect of Pressure on the Modulus of Rigidity of Several Metals and Glasses," *Jour. Appl. Phys.*, Vol. VIII (1937), pp. 129–34.

departure of the quantity $E\beta + 3(E/\mu)$ from 9 is less than 1.5 per cent over the 400°-temperature range. We regard this as an excellent agreement with the theoretical formula, in view of the fact that different methods, observers, and silica samples are involved. The margin of error is still large enough, however, to permit the computation from E/μ of a wave-velocity curve substantially different from that computed in terms of $E\beta$. One of these curves yields an average velocity-temperature coefficient almost twice that given by the other, although the actual values differ by less than 1.5 per cent. These discrepancies are pointed out in detail in order to make clear the extreme importance of having measurements of several elastic constants before drawing conclusions about the behavior of the wave-velocity with temperature or pressure. In Table 3 we also give for silica the variation with temperature of Poisson's ratio, σ, computed from the formula

$$\sigma = \tfrac{1}{2}\left(\frac{E}{\mu} - 2\right).$$

The curve of the wave-velocity in Pyrex with increasing temperature is similar to that in silica, both in magnitude and in shape. Minor irregularities in the curves seem to be characteristic of glasses, although they appear more strikingly in the compressibility than in the velocity curves.

The variation of wave-velocity in obsidian is small, less than 1 per cent in 300°. In soda-lime glass the wave-velocity decreases with temperature smoothly at the rate of 5.3 per cent in 300° C. In the Solenhofen limestone the decrease is about 10 per cent in 300° C. Near that temperature an irreversible change of some sort takes place, and the decrease in velocity becomes more gradual. The limestone will be discussed further in the section on rocks.

Our measurements on Pyrex and silica have an important bearing on the geophysical problem of the propagation of elastic waves in a substratum of hot glass. It is now apparent from direct experiment that short-period elastic vibrations may be supported by glasses of high silica content at atmospheric pressure and at temperatures up to at least 700° C. To this temperature we have substituted proof for reasonable supposition. It is also apparent that, if such glasses

can support vibrations at still higher temperatures, it will be because of the effect of high pressure on their elastic properties. The glassy substratum in the earth's crust, which has been proposed on geological evidence,[18] is presumed to lie at a depth of about 60 kilometers, where the temperature is probably at least 1,350° C. and the pressure 18,000 kg/cm². These conditions are not even approximated in our measurements. It is indeed difficult to see how elastic constants of rock and glass samples can ever be measured in the laboratory under combined pressures and temperatures of this magnitude. Since direct evidence is impossible to obtain with present methods, there is some excuse for speculation on the basis of the evidence we can obtain.

An important conclusion derivable from our measurements on Pyrex is that small alternating stresses of relatively high frequency may be supported by the glass at temperatures within the so-called "annealing range." The annealing of Pyrex glass has been studied by G. W. Morey at the Geophysical Laboratory of the Carnegie Institution.[19] He determined the rate of disappearance of strain by measuring the time required to reduce the birefringence in polarized light to 10 mμ/cm. This corresponds to an internal stress in the glass of about 3 kg/cm². Our measurements cannot be directly compared with those of Morey because the stresses in our experiments are much smaller in magnitude and in duration and because we measure quite different quantities. Some comparison is nevertheless possible. Morey shows that for Pyrex chemical resistance glass, the annealing constant, A, and the temperature, T, are related by the formula

$$\log A = 0.034\, T - 21{,}686\,.$$

Annealing times, for various temperatures, are inversely proportional to the annealing constants. He finds that it requires about two years of annealing at 430° C. to remove strains imparted to the glass. Only two hours are required at 550° C. If these data be ex-

[18] R. A. Daly, "Testing a Theory of the Earth's Interior," *Jour. Wash. Acad. Sci.*, Vol. XXV (1935), pp. 389–99.

[19] G. W. Morey, "Annealing of Pyrex Chemical Resistant Glass," *Ind. and Eng. Chem.*, Vol. XXVII (1935), pp. 966–71.

trapolated to 700° C. by means of the foregoing formula, it is found that at that temperature the annealing time is of the order of 0.07 seconds. The stresses which produce the vibrations in our sample alternate with a period of 0.00007 seconds. This is shorter by a factor of a thousand than the supposed annealing time. Our observation that Pyrex may be made to vibrate at 700° C. is thus entirely consistent with the probable annealing time as extrapolated from Morey's data. This confirms the expectation from the theory of annealing that stresses of short duration may be supported by the glass at temperatures where stresses of longer period would be neutralized by plastic flow. It would be interesting to perform further experiments with stress periods intermediate between the two hours employed by Morey and the 0.00007 seconds employed by us.

The curve of velocity of sound in Pyrex is also interesting because of its similarity in shape to the curve of rigidity in silica, as measured by F. Horton.[20] Horton obtained the modulus of rigidity in silica fibers as a function of temperature to 1,058° C., by measuring torsional vibration periods of a few seconds' duration. This modulus rises smoothly, with some downward curvature, to a maximum value at 876° C., and then turns downward with increasing speed as the temperature continues upward. Beyond the point of maximum rigidity the decrement of the vibrations begins to increase enormously, until resonant vibrations are no longer attainable above 1,058° C. The limiting factor in the attainable temperature is thus the increasing internal friction loss. This is the limiting factor also in our measurements on Pyrex, with the difference that the limitation appears at 700° C. in Pyrex. The temperature of maximum sound velocity (maximum Young's modulus) is 550° C. in Pyrex. The whole effect appears at about 350° lower temperature in Pyrex than in silica, but the general course of the curve is the same. It is highly probable that the velocity of sound in silica would be affected in the same way as its rigidity if our measurements were extended to higher temperatures. We thus conclude that the factor which limits the maximum temperature at which a glass will sustain vibrations is the internal friction loss. As this increases, the glass becomes aperiodic and can no longer be considered a rigid body. The limiting temperature un-

[20] *Op. cit.*

doubtedly increases with the silica content of the glass and with the frequency of the applied stresses. The effect of pressure on the limiting temperature is unknown but is of decisive importance to geophysical reasoning.

VELOCITY OF SOUND VERSUS TEMPERATURE IN ROCKS

The measurements of the velocity of sound in rocks as a function of temperature are graphically portrayed in Figures 4–9. The most

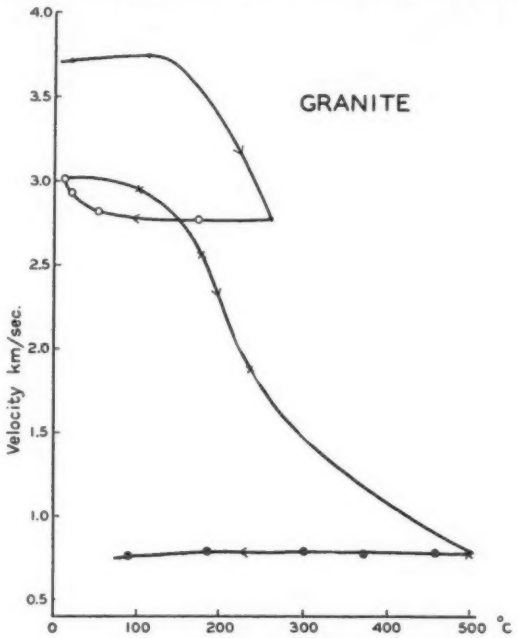

FIG. 4.—The velocity of sound as a function of temperature in Quincy granite

striking result is the irreversible character of the changes in velocity which take place on heating the material to successively higher temperatures. The velocity of sound decreases, in most cases very markedly, as the temperature rises, and does not increase again to its original value as the specimen cools. On heating the specimen a second time, the cooling curve is repeated until the temperature rises

beyond the maximum reached at the first heating. Then a further marked drop in velocity takes place. Again the velocity of the cooled sample fails to attain the value found at the beginning of the second

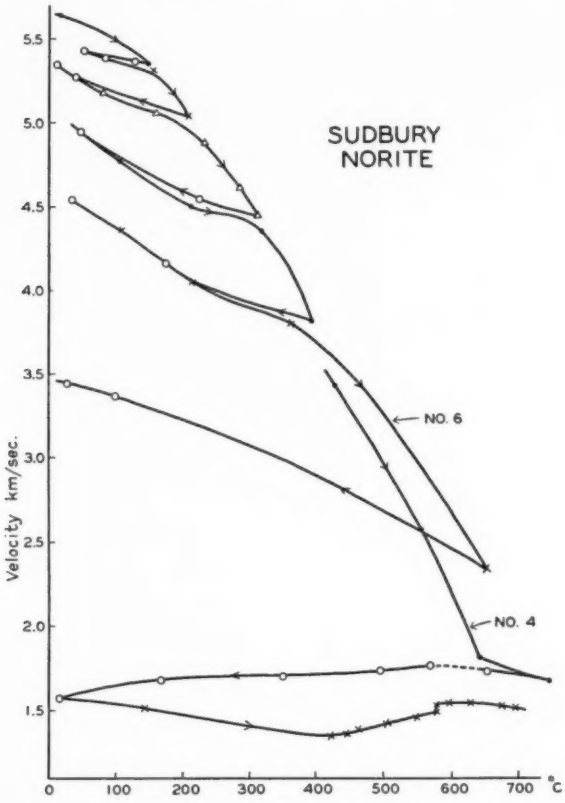

FIG. 5.—The velocity of sound as a function of temperature in two specimens of the Sudbury norite.

heating. On heating the material a third time to still higher maximum temperature, the same general pattern is repeated, until the value of velocity of sound is permanently reduced to a fraction of its value at the beginning of the experiment. In most cases the intensity

of the vibrations diminished somewhat at temperatures above 500° C. but was fully restored when the sample cooled. No effort was made to measure quantitatively the intensity or the decrement of the vibrations, but it was obvious that there was no change in these quantities comparable in magnitude with the changes observed in the velocity of sound.

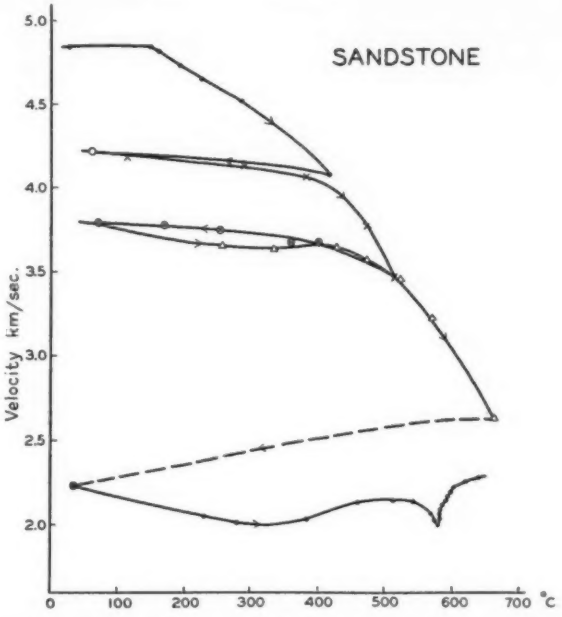

FIG. 6.—The velocity of sound as a function of temperature in quartzitic sandstone from Pennsylvania.

While no complete explanation of these extraordinary irreversible effects can be given without further experimentation, we suggest the hypothesis that they are caused by the thermal expansion of the crystals of which the rocks are composed. As the temperature of the sample rises, the crystals expand differently in different directions, depending on their composition and orientation. This unequal expansion may be expected to cause some internal cracking to take place along the lines of greatest stress. On cooling the specimen the

fractures would remain, the crystals would be more loosely joined, and the elasticity of the rock would be permanently reduced. If the sample were reheated, the crystals would expand again; but no further fracturing would take place until the temperature exceeded the maximum reached on the first heating. Above this temperature the cracking would continue and the elastic properties be still further impaired. Heating cycles to successively higher maxima might very probably produce velocity-temperature curves of the type which we have measured.

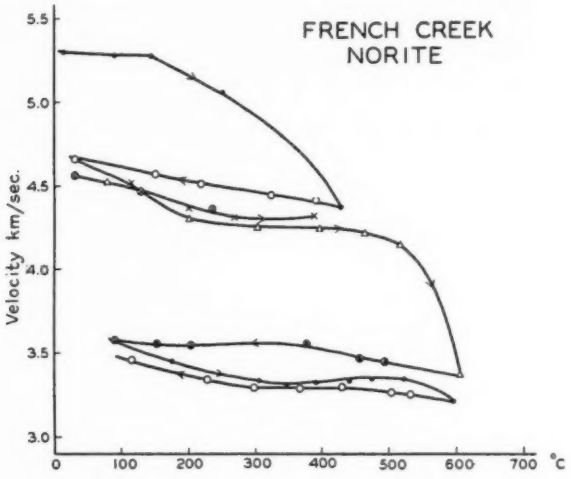

FIG. 7.—The velocity of sound as a function of temperature in French Creek norite.

In order to determine whether any visible alteration was produced in the rock structure by the heating cycles which we have described, thin sections of the diabase and Sudbury norite were examined before and after heating. In spite of the extraordinary reduction in measured elasticity, no change was detectable in the appearance of the thin sections. In both cases the sample looked slightly bleached, owing, presumably, to partial oxidation of iron compounds at the surface. The cylinder of Sudbury norite, after heating to 740° C., could be broken easily in the hands. We must therefore suppose that most of the breaking takes place along the crystal boundaries, so that

the individual mineral fragments are loosened and pushed apart. It seems curious that such marked mechanical weakening of the material could take place without altering the appearance of the thin section, but such was the case.

A similar hypothesis of internal fracturing due to the thermal expansion of the crystals was put forward by A. L. Day, R. B. Sosman, and J. C. Hostetter,[21] to account for the permanent increase in the specific volume of granite samples which had been heated above the

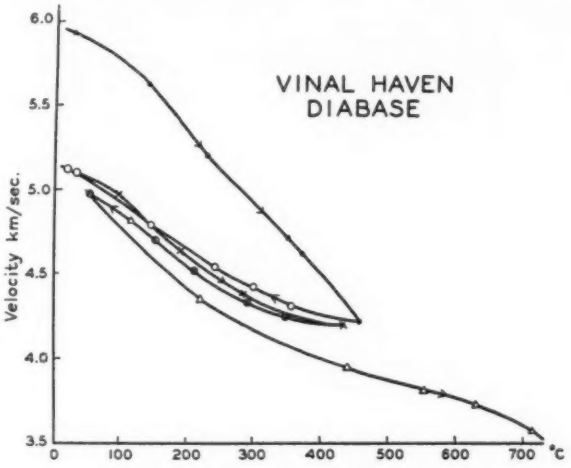

FIG. 8.—The velocity of sound as a function of temperature in Vinal Haven diabase.

inversion temperature of quartz. They found a volume increase, over the range from 0° C. to 575° C., of 5.07 per cent for one granite sample and 5.82 per cent for another. For pure quartz this is known to be 5.17 per cent. In each case most of the expansion took place in the hundred degrees before the inversion point was reached. They consequently attribute the shape of the granite expansion curve to the expansion of the pure quartz crystals pushing the other mineral fragments apart and increasing the apparent volume of the rock. While some of this expansion was reversible and disappeared as the

[21] A. L. Day, R. B. Sosman, and J. C. Hostetter, "The Determination of Mineral and Rock Densities at High Temperatures," *Amer. Jour. Sci.*, Vol. XXXVII (1914), pp. 1–39.

sample cooled, a certain amount of it was found to be permanent. Thus one of their specimens, heated twice to 600° C., showed a permanent volume expansion of 2.2 per cent; and one heated to 935° C. showed a permanent expansion of 9.3 per cent. Samples of granite heated to maximum temperatures lower than the quartz inversion point (575° C.) showed no measureable permanent increase in specific volume.

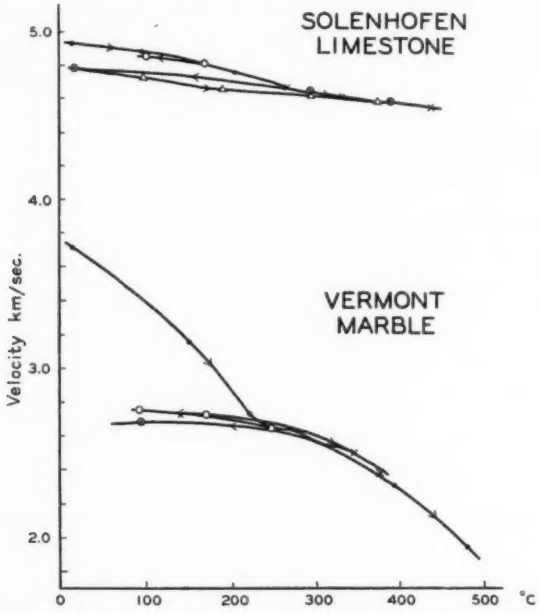

FIG. 9.—The velocity of sound as a function of temperature in Solenhofen limestone and in Vermont marble.

N. E. Wheeler[22] measured the thermal expansion of Westerly granite, Sudbury norite, and Carrara marble at temperatures up to 1,000° C. He found a permanent volume expansion in all these rocks—about 5 per cent in the granite, 2 per cent in the norite, and 1 per cent in the marble.

These measurements of thermal expansion in rocks are reviewed in some detail because they seem to be the only published work which

[22] N. E. Wheeler, *Trans. Roy. Soc. Canada*, Vol. IV, sec. 3 (1910), p. 19.

has any bearing on our measurements of the velocity of sound in rocks as a function of temperature. They prove that at least one other quantity, the specific volume, exhibits irreversible changes on heating. They also give evidence for the hypothesis that thermal expansion may cause a loosening of the rock structure which explains both the permanent volume expansion and the marked reduction in elasticity.

The only observed permanent changes in the specific volume of the samples studied in our experiments were in Vinal Haven diabase and Sudbury norite. The volume of the former increased about 1 per cent as the sample was heated to 710° C., and of the latter about 3 per cent as it was heated to 747° C. The volumes were determined from weight and dimensions and were not measured accurately enough to exhibit very small changes. The variations noted were, however, of the same order of magnitude as those observed by Wheeler and by Day, Sosman, and Hostetter.

Several of the samples, after heating, were soaked in water for several days and then weighed, to determine whether any marked increase in porosity had taken place. Since this test showed no difference between heated and unheated samples of the same rock, it was apparent that the elastic properties were the only ones drastically altered by the heat.

The effect of the quartz inversion appears definitely on two of the velocity-temperature curves, those for Sudbury norite No. 4 and for Pennsylvania quartzitic sandstone. The sandstone, containing 80 per cent of quartz, shows the effect more markedly than the norite, which contains only about 8 per cent of quartz. As the inversion, temperature, 575° C., is approached, the velocity in the sandstone drops suddenly, and then rises still more suddenly. The velocity in the norite shows only the sudden rise. The magnitude of the velocity "kink" is about 10 per cent in the sandstone and $2\frac{1}{2}$ per cent in the norite. The other sample of norite, No. 6, was not studied carefully in this range, and the granite sample was not heated above the inversion point, although these specimens should have shown the inversion "kink" if sufficient data had been taken. The other materials studied contained practically no quartz.

We should certainly expect some permanent impairment of elastic properties on heating a material so complex in structure as rock, but

the magnitude of the observed effect is surprising. For example, the sample of Quincy granite which we studied shows an irreversible reduction in velocity of sound from 3.72 km/sec to 0.78 km/sec on heating once to 500° C. This fivefold reduction in sound velocity represents a twenty-five–fold reduction in Young's modulus for this material. There was in this case no measurable change in specific volume, and the maximum temperature attained was well below the quartz inversion point. Thus the fracturing effect of thermal expansion must have been small in comparison with that produced in the samples studied by Day, Sosman, and Hostetter. Yet the reduction in elasticity is not only observable but is so large as to make the material completely unrecognizable as granite, so far as elastic properties are concerned.

The other rocks studied show a smaller reduction in velocity of sound than that which occurred in the Quincy granite. Sudbury norite No. 4, on being heated to 748° C., shows a permanent reduction in velocity of sound to one-fourth the original value. Sample No. 6 of the same material, heated to only 650° C., shows a velocity reduction from 5.7 km/sec to 3.45 km/sec. The velocity in French Creek norite dropped from 5.3 km/sec to 3.6 km/sec as the sample was heated twice to 600° C. The velocity in quartzitic sandstone from Pennsylvania dropped from 4.85 km/sec to 2.23 km/sec as the rock was heated to 667° C. The velocity in the sample of Vinal Haven diabase decreased from 5.93 km/sec to 5.1 km/sec on heating the material to 460° C. The velocity in the sample of white marble from Danby, Vermont, dropped from 3.72 km/sec to 2.7 km/sec on heating the material to 482° C. The sample which showed the smallest irreversible change in velocity of sound was the Solenhofen limestone. On heating this material to 213° C., there was no permanent velocity change. On reheating to 448° C. and cooling again, there was a permanent reduction in velocity from 4.91 km/sec to 4.79 km/sec, or about $2\frac{1}{2}$ per cent. It is possible that some of this change was caused by the driving-out of traces of organic matter from the limestone. The high porosity of this rock (about 6 per cent) may make possible considerable thermal expansion without internal fracturing.

The curves for limestone and marble, illustrated in Figure 9, are strikingly similar in character, although the magnitude of the

changes is much greater in marble than in limestone. In both cases there is no irreversible change until the temperature range 250°–300° C. is reached. As the temperature increases beyond this range to 480° C., little further alteration takes place. Since it occurs in both limestone and marble, this effect suggests that an abrupt structural change of some sort may take place in calcite between 250° and 300°. Further study of this might be of interest to mineralogists.

The outstanding, although unexpected result of the measurements on rock samples is accordingly the demonstration that the velocity of sound, and therefore Young's modulus of elasticity, is dependent to a very high degree on the structure of the substance. Changes in the physical structure which are too slight to make a measurable difference to the density or to the appearance of the thin section are capable of changing the modulus of extension by a large factor. The discovery of this property goes far toward explaining the discordant results frequently obtained in studies of the elasticity of rocks. A dynamic method is preferable to a static one, as was demonstrated in an earlier paper, since the structure-sensitive effects are minimized by the use of alternating stresses of small amplitude. Since even a dynamic method shows the large irreversible effects recorded here, it is regrettable, from the standpoint of measurement, that the earth's crust consists of such imperfectly elastic material as rocks.

High pressure must prevent the cracking to which we attribute the results described above. The observed increasing wave-velocities with depth in the earth's crust are utterly inconsistent with any such large-scale reduction of velocity as we have obtained in this research. The pressure of surrounding rock and the very gradual character of the temperature variations below the surface of the earth must prevent any loosening of the rock structures by unequal thermal expansion. There is ample time for plastic flow to relieve the stresses along crystal boundaries produced by changing temperature.

It is interesting to raise the question whether the other elastic constants of rocks would also show irreversible effects if they were measured as a function of temperature under laboratory conditions. This question is partially answered by comparing our measurements on the Solenhofen limestone with compressibility measurements of Birch and Dow on the same material. Table 2 contains these data.

On heating to 448° C., Young's modulus, computed from the velocity of sound, shows an irreversible reduction of 5 per cent. The compressibility, on heating to 476° C., shows an irreversible reduction of 10 per cent. The computed wave velocity in an infinite medium shows an irreversible increase of 3 per cent. If the behavior of this limestone is representative, we conclude that heating at atmospheric pressure subjects both Young's modulus and the compressibility to permanent alterations of the same order of magnitude. Further measurements of the compressibility of rock samples before and after heating are needed to confirm this conclusion.

The original object of this research was to measure temperature coefficients of velocity of sound for these rocks. This proved to be difficult, since we have no dependable means of separating the reversible temperature effect from the relatively large irreversible effect due to internal fracturing. We may, nevertheless, suggest approximate temperature coefficients based on the slopes of the velocity curves below 400° C. In cases where the samples were subjected to several successive heatings, the low-temperature portions of all the heating and cooling curves are reasonably parallel. This enables us to estimate at least the order of magnitude of the temperature coefficient. It is probable that the actual temperature coefficient of velocity of sound in the granite sample is less than ± 1 per cent per hundred degrees; in the Vinal Haven diabase, about -5 per cent per hundred degrees; in the Sudbury norite, about -4 per cent per hundred degrees; in the French Creek norite, about -1.6 per cent per hundred degrees; in the quartzitic sandstone, about -1 per cent per hundred degrees; in the Solenhofen limestone, about -1.2 per cent per hundred degrees; and in the Danby marble, about -3 per cent per hundred degrees. These values may be compared with -1.34 per cent per hundred degrees for steel.

As a basis for speculation it seems plausible to assume that the effects of temperature and of pressure on the wave-velocities in rocks may be added together to give the approximate combined effect. A reasonable estimate of the temperature at a depth of 10 kilometers is 300° C. The results described here for granite indicate that the total temperature effect on the wave-velocity in the granitic portion of the crust is probably less than 2 per cent. The bulk of the observed increase in longitudinal wave-velocity in the upper layer

must be attributed to the effect of pressure increasing with the depth. In view of the much higher temperature coefficient of velocity observed by us in two types of diabase (Vinal Haven and Sudbury), we suggest that increasing temperature may play a much larger role in determining the velocity of elastic waves in diabase than in granite. If diabase actually possesses a negative wave-velocity temperature coefficient of 5 per cent per hundred degrees, the gaps in our knowledge of this substance become embarrassingly large. In the absence of any information about the effect of pressure on Young's modulus of diabase, we are at liberty to assume a net *decrease* in this constant with depth below the surface of the earth. In order to compute a curve of wave-velocity with depth from their compressibility measurements, Birch and Dow make the assumption that Poisson's ratio remains unchanged under heat and pressure. This requires that Young's modulus increase 40 per cent in the 40-kilometer depth range. A decrease in this constant seems quite as reasonable in view of the temperature effects just described. The result is a large margin of uncertainty, which can be resolved only by the measurement of the rigidity or of Young's modulus in diabase under combined heat and pressure.

Summarizing our results, we find that the metals and glasses give smooth and repeatable curves, from which we are able to compute longitudinal wave-velocities as a function of temperature, at atmospheric pressure. This is done by combining our measurements with the compressibility measurements of Birch and Dow on the same materials. Further, we obtained interesting information on the upper temperature limit at which glass will support vibrations, showing that this limit is 700° C. in Pyrex and near 1,000° C. in silica, and that the factor which causes the glass to cease transmitting vibrations is the increasing decrement or internal friction loss. The results on rocks were complicated by their imperfect elastic behavior but permit the suggestion that the effect of temperature on granite is small enough to be neglected in comparison with the effect of pressure, while in diabase the effect of temperature may be of decisive importance. Since diabase is known to be a widespread constituent of the rocky crust, particularly at the depths where the temperature is high, our results emphasize the importance of further work on this material.

Acknowledgment:—The author wishes to acknowledge his indebtedness to Dr. Francis Birch for the construction of the furnaces and thermostat, and to the Harvard Committee on Geophysical Research for financial assistance in the purchase of apparatus. It is a pleasure to acknowledge also the continued interest and counsel of Professor R. A. Daly, of the Harvard department of geology.

APPENDIX

THE DAMPING OF VIBRATIONS IN ROCKS

The decrements of the samples are frequently referred to in the text. Since nothing on this topic with specific reference to rocks has been published, it seems advisable to discuss it briefly here.

The rate at which the amplitudes of elastic waves in solid bodies subside with increasing time and distance should be of interest in geophysical problems. The damping or decrement of elastic vibrations gives a measure of the rate at which cyclic strains are dissipated in internal friction. The damping in various types of rock is measured in these experiments in terms of the sharpness of resonance. If f_1 and f_2 are the frequencies on either side of resonance at which the amplitude of vibration of the sample falls to $1/\sqrt{2}$ times the amplitude at the resonant frequency, f_0, the dissipation constant, Q, of the material is given by the relation

$$Q = \frac{\text{Elastic reactance}}{\text{Resistance}} = \frac{f_0}{f_1 - f_2}.$$

This quantity, from our measurements, is tabulated below for representative rocks. Values for quartz, brass, glass, and lead are quoted for comparison from the work of R. E. Wegel and H. Walther. Their paper on internal dissipation in solids, published in *Physics*, Vol. VI (1935), p. 141, should be consulted for further discussion of the subject.

Material	Q
Vinal Haven diabase	600
Dolomite (Pa.)	390
Sudbury norite	290
Slate (Pa.)	265
French Creek norite	170
Limestone (Pa.)	150
Quartzitic sandstone (Pa.)	130
Granite (Rockport, Mass.)	130
Granite (Quincy, Mass.)	100
Gneiss (Pelham, Mass.)	55
Marble (Vt.)	20
Quartz	100,000
Brass	40,000
Glass	1,600
Lead	50

PETROLOGY, STRUCTURE, AND RELATION TO TECTONICS OF PORPHYRY INTRUSIONS IN THE BEARTOOTH MOUNTAINS, MONTANA[1]

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ABSTRACT

Many of the porphyry intrusions of the northeastern Beartooth Mountains are in close proximity to the Beartooth thrust fault which is thought to have been developed during the Laramide orogeny. Detailed studies of the porphyry bodies show that they were intruded before the thrusting took place and orogenically have no relation to the compressional episodes expressed by the thrusting. The shape, structural and age relations, distribution, and mineral composition of the different porphyry intrusives are described.

INTRODUCTION

The purpose of this study has been to determine the relation of the tectonics and intrusion of the porphyries along the northeastern side of the Beartooth Mountains and particularly to determine whether or not the porphyry intrusions had any discoverable relation

[1] The regional study was undertaken by John T. Rouse and H. H. Hess during the summer of 1933 as part of the Yellowstone-Bighorn co-operative research program in Montana and Wyoming. Detailed studies concerning the porphyries in the McLeod area, the Nye quadrangle, and the Red Lodge quadrangle were made by Freeman Foote, John S. Vhay, and K. P. Wilson, respectively, in connection with their studies of these specific regions. The authors wish to thank Professors W. T. Thom, Jr., and R. T. Chamberlin for numerous profitable field discussions and helpful suggestions in the preparation of the manuscript. We are also indebted to Professors Erling Dorf and Elwyn Perry for information concerning the Paleozoic section along the Beartooth front at Red Lodge, Montana. The writers gratefully acknowledge the chemical analyses which were furnished by the department of geology of Princeton University.

to the thrusting[2] around the edge of the Beartooth Mountains. Since the porphyry intrusions are in close proximity to the thrust faults some have suggested that the porphyry magmas might have "greased" the thrust planes and so facilitated movement along those planes. An attempt has been made to determine the age of the porphyries in order to correlate them with the tectonic conditions at the

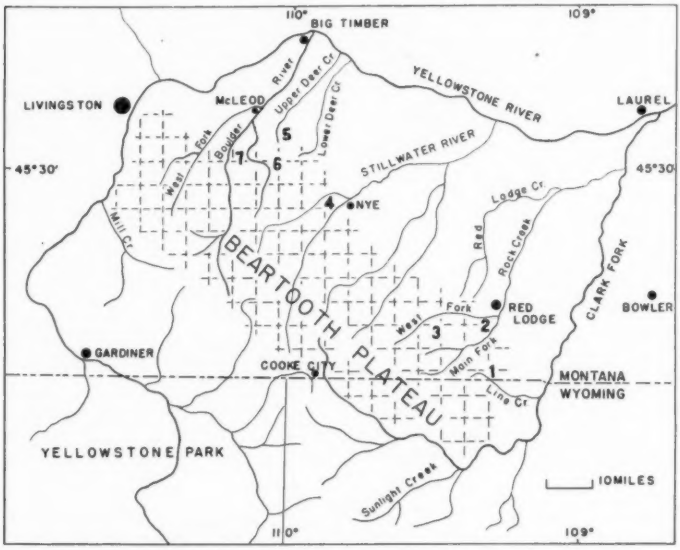

FIG. 1.—Sketch map of the Beartooth Mountains area showing location of porphyry bodies described in this paper. Numbers refer to topic headings in the text. *1*, Line Creek; *2*, Red Lodge Front; *3*, Beartooth Plateau; *4*, Nye Quadrangle; *5*, Gold Hill; *6*, Lodgepole Intrusive; *7*, East Boulder Plateau.

time of intrusion. A study of the outward shape and internal structures of the porphyry bodies also supplied some direct evidence on the conditions under which they were intruded. In some cases consanguinity between the porphyries themselves or between the porphyries and the volcanics of a given area was recognized. The sketch map, Figure 1, shows the general location of the area as well as the specific regions studied.

[2] W. H. Bucher, R. T. Chamberlin, and W. T. Thom, Jr., "Results of Structural Research Work in Beartooth-Bighorn Region, Montana and Wyoming," *Amer. Assoc. Pet. Geol. Bull. 27* (1933), pp. 681–93.

DISTRIBUTION, OCCURRENCE, AND PETROGRAPHIC DESCRIPTION OF THE PORPHYRIES

In a broad way the porphyries under consideration are confined to the northeast portion of the Beartooth Mountains where they occur as:

1. Small intrusive sills, principally in the Cambrian sediments, cropping out along the border of the mountain range where the present erosion surface intersects the Cambrian formations.

2. Nearly horizontal sheetlike masses on the Beartooth Plateau, at or not far below the Cambrian–pre-Cambrian contact.

3. A concentration of intrusives, in many places laccolithic, where the Nye-Bowler lineament[3] approaches and parallels the mountain front (Nye to McLeod).

At the same time it should be noted that the porphyries are absent in the mountains east of the Beartooths, namely, the Pryor and Bighorn mountains.

RED LODGE FRONT

A continuous sheetlike body of gray porphyry intrudes the nearly vertical Gros Ventre (Cambrian) shale along the Beartooth front approximately 3 miles southwest of Red Lodge, Montana. Various parts of this intrusive have been displaced by several transcurrent faults or flaws[4] developed at the time of overthrusting (Fig. 2).

Megascopically the porphyry is composed of white, idiomorphic feldspar phenocrysts $\frac{1}{4}$ inch in diameter, imbedded in a gray groundmass which is microcrystalline near the margins of the sheet and finely crystalline near the center. Occasional rounded quartz phenocrysts are also found but ferromagnesian phenocrysts are quite scarce. These porphyries are very similar to those found farther west along the mountain front in the Stillwater region.

In thin section the groundmass is exceedingly fine grained and the majority of its mineral constitutents are too small for determination. The dominant phenocrysts are albite ($Ab_{90}An_{10}$) which is, in many cases, beautifully zoned; and practically all the albite is badly fractured and altered. Altered and fresh crystals of hornblende are

[3] Charles W. Wilson, Jr., "Geology of the Nye-Bowler Lineament, Stillwater and Carbon Counties, Montana," *Amer. Assoc. Pet. Geol. Bull. 20* (1936), pp. 1161–88.

[4] Elwyn L. Perry, "Flaws and Tear Faults," *Amer. Jour. Sci.*, Vol. XXIX (1935), pp. 112–24.

common as well as smaller quantities of biotite and rounded quartz. Small grains of magnetite and needles of apatite are scattered throughout the groundmass. This porphyry is quartz monzonite having the following modal composition: quartz, 3 per cent; ortho-

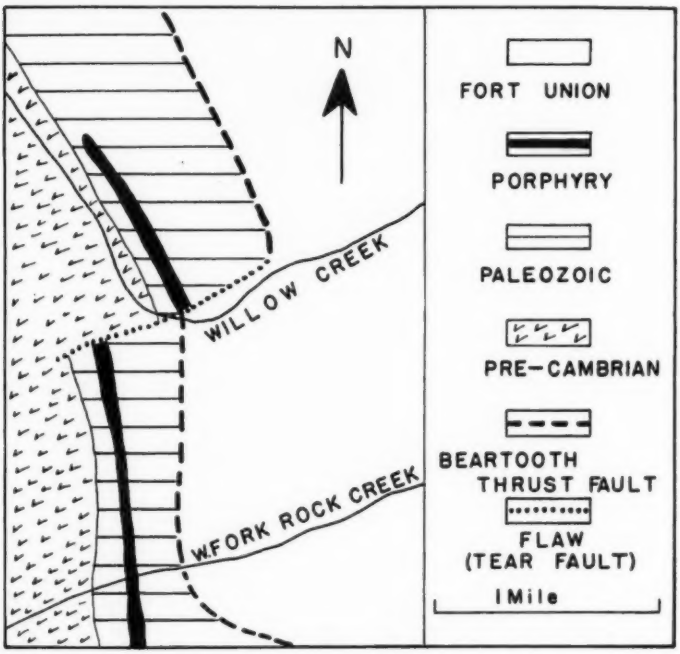

FIG. 2.—Sketch map of the West Fork–Willow Creek area showing the porphyry body offset by transcurrent or tear faults (modified after Perry).

clase, 6.2 per cent; albite, 21.4 per cent; biotite, 4.8 per cent; hornblende, 3.6 per cent; groundmass, 61 per cent.

LINE CREEK

On Line Creek (Wyoming-Montana state line), a tributary of the Clark Fork, a poorly exposed porphyry sheet is intrusive into the Cambrian sediments. This porphyry is identical, both megascopically and microscopically, with that just described in the Red Lodge

area. Structurally the relations are also similar in that on Line Creek the porphyry has been intruded into the incompetent Cambrian shales.

LODGEPOLE INTRUSIVE[5]

A series of porphyry intrusions occurs a few miles north of the Beartooth Front between the Stillwater and Boulder rivers (McLeod area). Many of these intrusives are of the laccolithic type and show rough parallelism with a line of domed or anticlinal structures extending from Livingston to the Pryor Mountains and called the Nye-Bowler lineament by Wilson.[6] The larger intrusive bodies are quartz diorites and the smaller ones range from quartz latites to dacite porphyries.

The largest mass of porphyry exposed in this area, the Lodgepole intrusive (referred to by Wilson as the intrusion at Squaw Peak), is situated at the headwaters of Deer Creek between the drainage systems of the Boulder and Stillwater rivers (Fig. 1, location *6*). Part of this body has been mapped as acid porphyrite by Iddings and Weed.[7] The general form of the intrusive is that of a Christmas-tree laccolith whose marginal sills are now steeply dipping because of the later injection of a large central mass of porphyry.

The central portion of the body is holocrystalline, fine-grained hornblende quartz-diorite porphyry. The border facies is a dacite which has a gray aphanitic groundmass and which contains phenocrysts of quartz, feldspar, biotite, and hornblende. Thin sheets of dacite, similar to the chilled facies of the quartz diorite, occur between the bedding planes of the lower Paleozoic roof rock above the main mass of the intrusive. Xenoliths are numerous in the quartz diorite, the more abundant ones being of altered norites and ultramafic rocks evidently carried up from the underlying pre-Cambrian Stillwater complex.[8] Xenoliths of pre-Cambrian schists and Paleozoic limestones are also common.

[5] The structure and petrology of this intrusive will be discussed more fully by Foote in a paper now in preparation.

[6] *Loc. cit.*

[7] "Livingston Folio," *U.S. Geol. Surv., Folio 1* (1894).

[8] A. L. Howland, J. W. Peoples, and Edward Sampson, "The Stillwater Igneous Complex and Associated Occurrences of Nickel and Platinum Group Metals," *Mont. Bur. Mines and Geol. Misc. Contrib. 7* (1936).

In thin section the dacite of the border facies shows a groundmass of a finely crystalline aggregate of quartz and feldspar with considerable quantities of exceedingly fine-grained magnetite. The phenocrysts consist of a few large rounded and embayed grains of quartz, orthoclase, andesine, common hornblende, and a few large crystals of biotite. Magnetite, apatite, and carbonate occur as common accessories. Considerable alteration of the feldspars to sericite and kaolin has taken place, some of the grains being almost entirely altered while in others sericitization has taken place only along the crystal boundaries. In places the biotite shows patchy alteration to leucoxene.

The central quartz diorite is essentially the same mineralogically as the border facies, the main difference being that quartz is absent as phenocrysts but does occur sparingly as small grains in the groundmass. Phenocrysts of orthoclase, andesine, common hornblende, and biotite form the major portion of the rock. Some of the andesine shows well-developed zoning. Alteration of the feldspar is not so pronounced as in the dacite, being confined mostly to sericitization along the borders.

The age relations between the porphyries and the Beartooth thrusting have been clearly established through detailed work on this intrusive. The Lodgepole intrusive is situated north of the Beartooth thrust which, in this vicinity, dips to the north causing duplication of the Paleozoic and Mesozoic beds. Some of the sills in the Gros Ventre (Cambrian) shales associated with the intrusive are very widespread and are found south of the Beartooth thrust. At one locality at the head of Blacktail Canyon, approximately 2 miles south of the thrust, northward dipping thrusts are found to cut the porphyry sills, which is taken as evidence that the sills are older than the thrusting.

The general shape of the intrusive is circular and domelike although the floor is now tilted. This would seem to preclude the possibility that the intrusion took place at the same time as the thrusting, for in that case the sediments should have been folded into an elongated anticline with its long axis parallel to the thrust plane, or if a laccolith were intruded it would be asymmetrical. Mesozoic beds on the northern side of the intrusive have been upturned by the

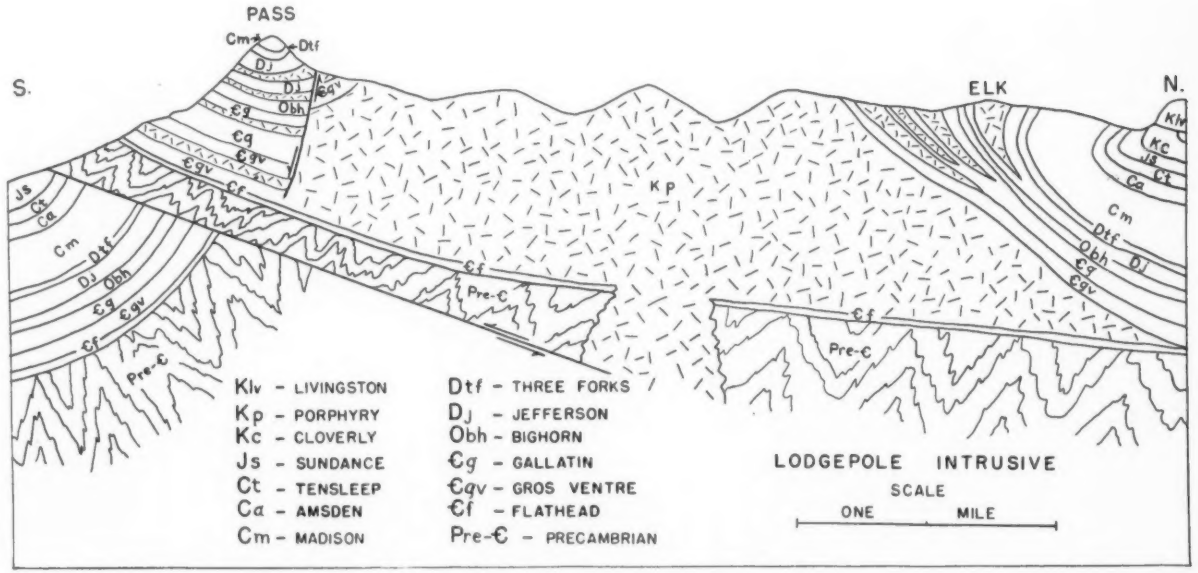

FIG. 3.—Cross section of Lodgepole intrusive (Freeman Foote)

doming action of the porphyry and are overlain with angular unconformity by the Livingston volcanics, here considered to be equivalent in age to the upper part of the Montana group. A cross section of this intrusive is shown in Figure 3.

ENOS MOUNTAIN

Enos Mountain, approximately 3 miles from the East Boulder Front, between Elk and Enos creeks, is a breccia plug. This intrusion differs from the others in that it broke through to the surface and probably formed one of the vents from which the adjacent volcanic breccias were extruded. The material in the vent is an autobreccia of dacite fragments (1–2 inches in diameter), together with occasional limestone fragments in a dacite matrix. Limestone fragments are not confined to the vent but are also present in the volcanic breccia which rests on the nearby surface. Zones of brecciation were noted in the Madison (Mississippian) limestone which forms the eastern edge of the vent.

In the sedimentary rocks, which are upturned and even overturned around the vent, sheets of non-brecciated dacite are found. These, similar to the sheets of the Lodgepole intrusive, were probably intruded first and then later overturned along with the sediments as the main mass of the intrusive welled up and later broke through to the surface in a volcanic eruption. The section beneath the breccias and away from the vent was tilted by the thrusting and, as the breccias themselves conform to this dip, they too must have been tilted at the same time and are, therefore, prethrusting in age. Breccias of similar composition (dacitic) to those in the vent are found a short distance west of the vent where they overlie a sandstone which is probably the Eagle (Upper Cretaceous). On top of these breccias is a great thickness of the Livingston volcanic tuffs and agglomerates which are mostly andesitic in composition. In the direction of Columbus to the north the Livingston series interfingers with the Lance, so it seems that volcanic activity, though chemically of a different type, continued for some little time after the porphyry intrusions. The vents from which the Livingston volcanics were erupted are not known but cannot be far from the mountain front as the series apparently thins toward the north.

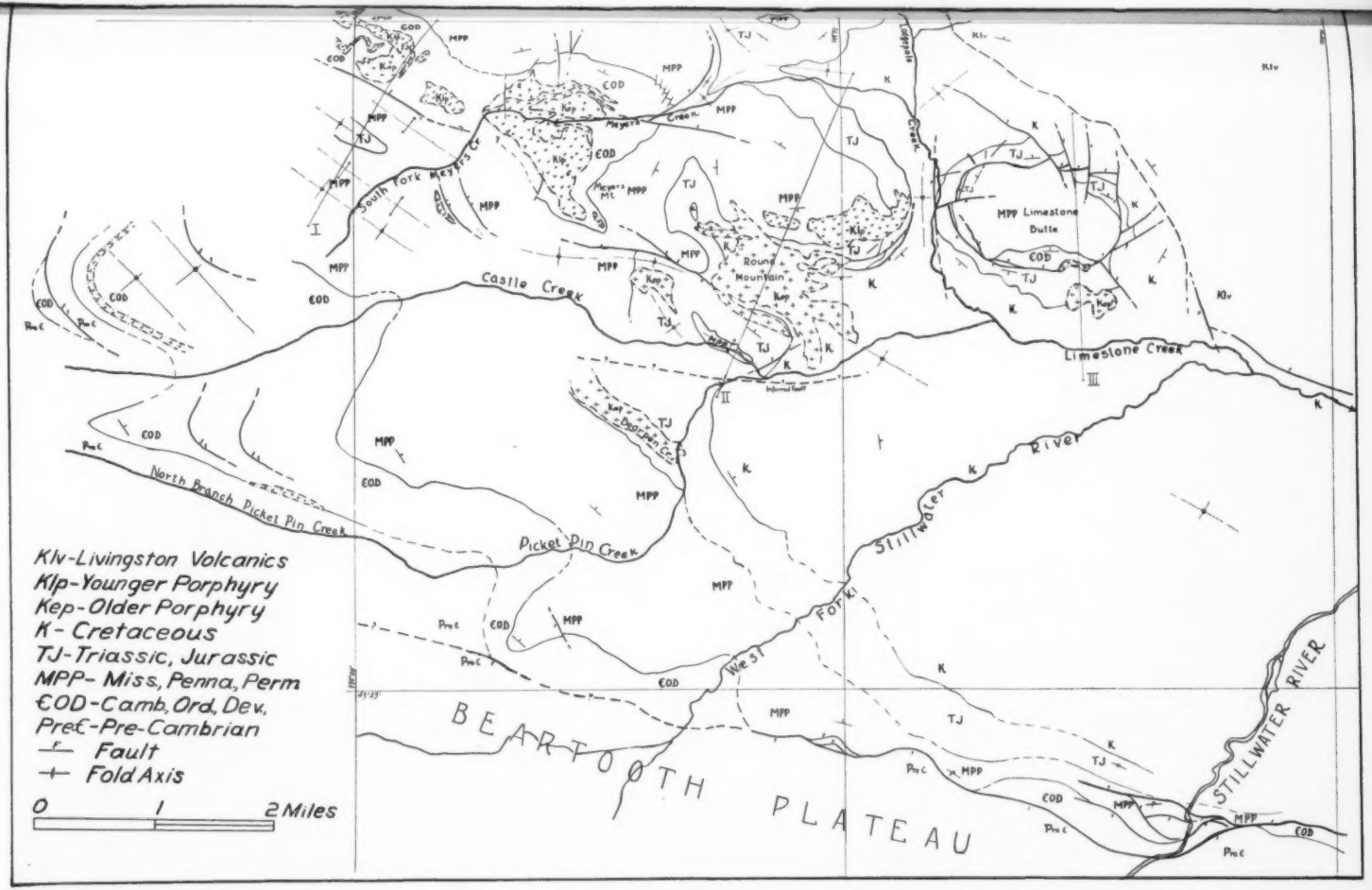

FIG. 4.—The general geology of the northwest portion of the Nye Quadrangle (taken from an unpublished Ph.D. thesis, "The Geology of a Part of the Beartooth Mountain Front near Nye, Montana," by J. S. Vhay [1933]).

GOLD HILL

Gold Hill, about 6 miles northeast of the Lodgepole Intrusive, (Fig. 1, location *5*) is probably a laccolith about which the Paleozoic rocks outcrop in a ringlike form. This porphyry intrusive has been greatly altered by hydrothermal solutions which brought in ore minerals. Its position, several miles from the mountain front and the present outcrop of the Beartooth thrust, is interesting in that it indicates the lack of dependence of this group of porphyry intrusives on the Beartooth Mountain thrusting.

NYE QUADRANGLE

The porphyry intrusions of the Nye Quadrangle are of two general forms: rather large complex laccolithic bodies and thick sills (for general location see Fig. 1, location *4;* and for a detailed map of this area see Fig. 4.)

The group of laccolithic bodies lies along the westward extension of the Nye-Bowler lineament which is believed to represent a major line of weakness. At Limestone Butte the porphyry body takes the form of a faulted laccolith (bysmalith). Here the sharp updoming (Fig. 5, III) of the sediments was accompanied by the formation of a ring fault which has the greatest displacement on the south side of the butte. In addition numerous radial faults were formed (Fig. 4). On the north end of a syncline in the vicinity of Round Mountain there is a complex laccolithic intrusion of quartz keratophyre. Around the margins of this laccolith many sills extend out into the less resistant shales (Fig. 5, II). On the ridge northeast of Round Mountain a somewhat concordant mass of dacite intrudes the nose and flank of an elongated dome which itself may be the result of an intrusion. In Meyers Canyon a compound intrusive body of quartz keratophyre and dacite arches the entire Paleozoic section and shows cross-cutting relations as high as the Devonian. In North Meyers Canyon two quartz keratophyre masses dome the sediments and a number of dacite sills intrude the Cambrian.

South of the laccoliths are sills which seem to be limited to the less resistant shales. On Bearpen Creek a thick sill of quartz keratophyre intrudes the Triassic shales, and another porphyry is found in the shales and thin-bedded limestones beneath the Pilgrim limestone (Cambrian) on the ridges north and south of Castle Creek.

In general the porphyries are of two kinds: quartz keratophyre and dacite. These may occur together in the same body and where they do the relations are such that the dacite appears to be younger.

The quartz keratophyre is fine grained and white to cream in color. The phenocrysts are not abundant and are rather small. In

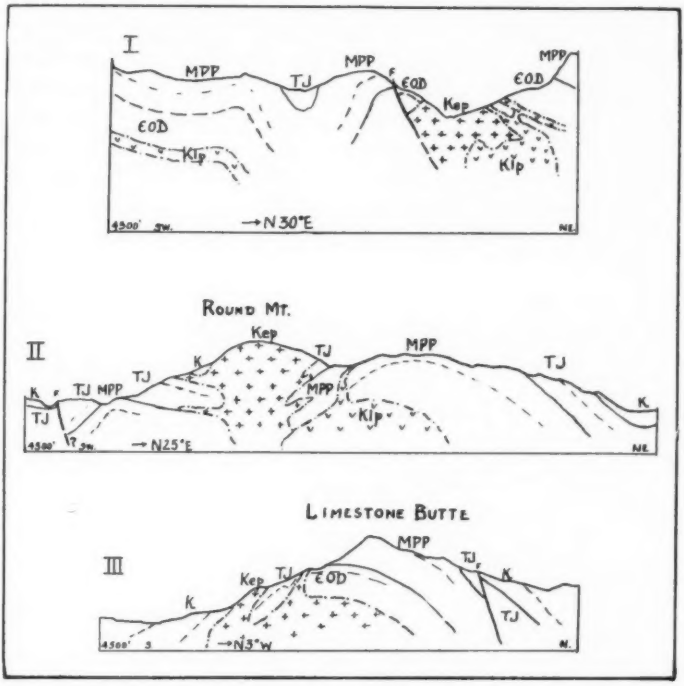

FIG. 5.—Cross sections through some of the intrusives of the Nye Quadrangle. Roman numerals refer to the location of the sections on the map (Fig. 4). Legend for cross sections same as map legend. (Source of sections same as Fig. 4.)

thin sections they are seen both as clusters of crystals and as separate individuals which are mostly of sodic oligoclase and orthoclase, with the former in greater abundance, together with scattered biotite phenocrysts. The fine-grained matrix consists of oligoclase laths, in places showing trachytic structure; irregular, rather equidimensional spots of orthoclase (?); considerable interstitial quartz; biotite flakes;

and fine specks of magnetite. In places some sericite and carbonate have replaced the feldspars. The chemical analysis and computed norm of the quartz keratophyre are given in Table 1.

The dacite, light to gray in color, is of coarser appearance than the quartz keratophyre and has larger and considerably more numerous phenocrysts of feldspar and biotite. In thin sections the feldspar phenocrysts are seen to be almost completely replaced by carbonate and sericite, but the biotite is usually unaltered. Numerous areas of chlorite and sericite, outlined by regular lines of magnetite, strongly suggest completely replaced crystals of hornblende. Small phenocrysts of quartz may be present in some sections while in others rather large, tabular resorbed crystals may be seen. The matrix is similar to that of the quartz keratophyre but is coarser and more apt to show trachytic structure. The dacite has the same arrangement of plagioclase laths in and around irregular areas of orthoclase and interstitial quartz. A chemical analysis and norm are given in Table 1.

The shape of the intrusives and their relation to the structures in the adjacent country rock in the Nye area indicate that the porphyries were intruded prior to the thrust faulting which occurred late in the Laramide orogeny. The synclines in the sediments around and between the laccoliths have been compressed and overthrust by the domes to a greater extent than would seem likely if they were merely complementary to the domes. Additional deformation took place during the Laramide period of compression and the porphyry bodies acted as resistant masses between which the synclines were mashed. At the same time the whole area was underthrust from the southwest so that asymmetrical folds and thrust faults, dipping northeast, were formed. These are especially well shown on the ridges on each side of the upper part of Castle Creek. With respect to the plains area the movement of the entire northeast side of the Beartooth Mountain block appears to have been in a direction a little north of east. East of the Nye area the pre-Cambrian is thrust over the younger rocks. On the southwest side of the general area intruded by the porphyries, however, the mountain block underthrusts to the northeast. The reason for this may be, in part, that because of the mass of the porphyry the area stood higher and had greater competence than the surrounding area.

Whatever may be the explanation of the thrust relations mentioned above, it seems most probable that the porphyries were already emplaced when the major deformation occurred. Further evidence of this lies in the suggestion of a local unconformity at the base of the Livingston formation in the vicinity of Limestone Butte. Elsewhere the base of the Livingston volcanics, which is thought to mark the initiation of part of the Laramide disturbance, interfingers with the Judith River.[9] This local unconformity suggests that the Limestone Butte structure had been formed and exposed to erosion before the beginning of Livingston time.

EAST BOULDER PLATEAU

A large body of porphyry, approximately 750 feet wide and 1,500 feet long, has been intruded parallel to the banding of the Stillwater gabbro (pre-Cambrian) near the mountain front at the north end of the East Boulder Plateau (Fig. 1, location 7). About one-half of the total bulk of the porphyry consists of phenocrysts, the majority of which are feldspar. A few rounded quartz grains and some biotite are the other phenocrysts.

In thin sections the euhedral feldspar phenocrysts (3–6 mm.) are the most conspicuous. They are badly altered and are probably oligoclase. The quartz grains, 5 mm. in diameter, are rounded and resorbed and the greenish brown biotite is rimmed by small magnetite octahedrons. A few crystals of zoisite are present and apatite is fairly abundant as an accessory. Secondary minerals replace a ferromagnesian mineral now completely removed. Its former presence is indicated by pseudomorphs which consist of small green chlorite aggregates with calcite and some apatite present around and within the pseudomorphs. Aggregates of calcite and small pyrite crystals either fill miarolitic cavities or replace small areas of the rock. All of these secondary minerals are probably deuteric. The rock is a quartz-monzonite porphyry.

BEARTOOTH PLATEAU

The porphyry intrusions within the Beartooth Mountains, although their outcrops are not closely associated with the thrust faults along the mountain front, merit consideration because of their

[9] Wilson, *op. cit.*

close mineralogic relationships and structural similarities to the porphyries just described.

Where erosion has not removed the Cambrian completely in the mountainous areas, small narrow sills of rather fine-grained porphyry are found in the Gros Ventre shales. These are predominantly dark gray in color and contain phenocrysts of glassy feldspar as large as $\frac{1}{4}$ inch in diameter. The groundmass is generally aphanitic.

Larger porphyry bodies occur south of the Paleozoic mountain front in the pre-Cambrian crystalline rocks and assume various shapes from dikes through sheets to bodies which are laccolithic in shape. These will be considered under four separate headings:

1. The principal body, laccolithic in shape, is on the Line Creek Plateau and appears to have a slight pitch to the southwest. Its maximum visible thickness of 1,000 feet is reached along the canyon wall of the Main Fork of Rock Creek, but it shelves rapidly toward the northeast where the contact emerges from the plateau. The basal chilled zone of 100 feet is very dense and glassy but the groundmass becomes progressively more granitic as the central portion of the mass is approached.

2. The majority of the intrusions in this region are in the form of sheets dipping 30°–40° and varying from 50 to 500 feet in thickness.

3. The few dikes that are found have a rambling habit and work around obstructions rather than cut across them.

4. One conical body of small dimensions was found, at the head of Red Lodge Creek, that increases rapidly in diameter from the bottom of the canyon upward. There is a very marked shearing and crushing of the granite gneiss on the contacts, and the chilled zones of the porphyry itself show crushing of the plagioclase to thin plates, while the biotite plates are arranged in a fluxion structure. The orientation of these plates is only roughly parallel to the fluted contact and their edges often emerge at the sides of the fluting.

Most of the porphyries are gray but brown ones are not uncommon. Mineralogically the gray porphyries contain, as phenocrysts, plagioclase, orthoclase, biotite, and some quartz. The plagioclase, somewhat altered around the edges but clear in the center, is well zoned and these zones range, from the center outward, from approximately An_{30} to An_{10}. Cracks in some of the plagioclase may be filled

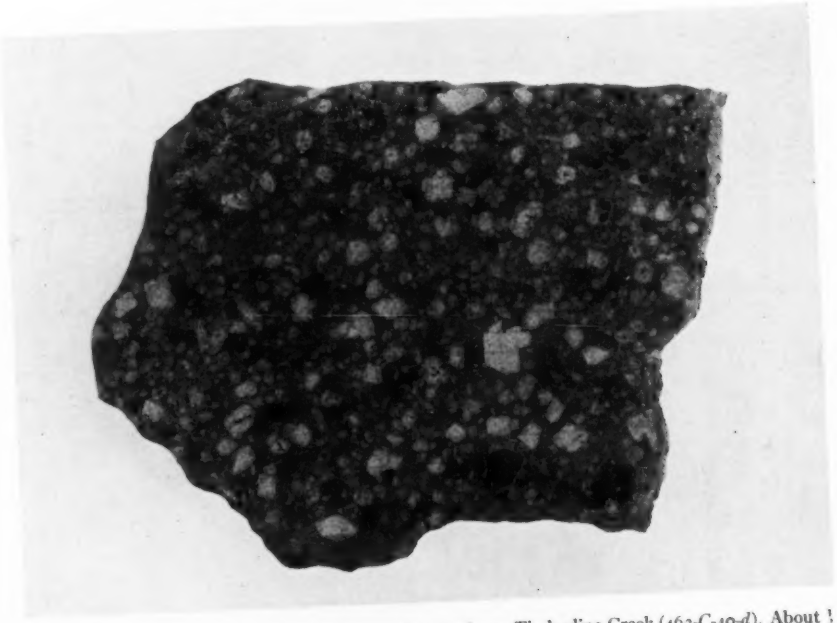

FIG. 6.—Photograph of gray quartz monzonite porphyry, Timberline Creek (462-*C*-40-*d*). About ½ natural size.

TABLE 1

CHEMICAL ANALYSES AND NORMS OF THE PORPHYRIES

	CHEMICAL ANALYSES				
	I	II	III	IV	V
SiO_2	62.27	65.94	65.74	70.82	66.11
Al_2O_3	17.20	16.79	16.49	15.45	15.44
Fe_2O_3	0.88	1.50	0.66	0.58	1.30
FeO	0.49	0.25	1.01	0.65	1.15
MgO	0.47	0.22	0.83	0.48	0.97
CaO	1.91	2.25	3.33	1.53	2.91
Na_2O	6.07	5.61	4.37	4.52	3.76
K_2O	3.35	4.47	2.35	3.66	3.01
H_2O+	0.63	0.57	1.56	0.99	1.88
H_2O-	0.11	0.22	0.56	0.46	0.47
CO_2	0.91	1.65	2.64	0.55	1.74
TiO_2	0.19	0.20	0.17	0.12	0.27
P_2O_5	0.13	0.06	0.09	0.04	0.67
MnO	0.04	0.09	0.07	0.05	0.06
BaO	0.09	.....	.....	0.22	.....
Total	99.74	99.82	99.87	100.12	99.74
	NORMS				
Quartz	16.56	15.54	29.76	27.72	30.24
Orthoclase	20.02	26.69	13.90	21.65	17.79
Albite	51.35*	47.16†	36.68‡	38.20§	31.96‖
Anorthite	3.06*	0.83†	0.00‡	4.26§	4.17‖
Ilmenite	0.30	0.46	0.30	0.18	0.46
Magnetite	1.16	0.46	0.93	0.84	1.86
Hematite	.....	1.12	.....	.....	.....
Corundum	2.44	2.45	6.83	2.47	4.39
Hypersthene	($MgSiO_3$, 1.70; $FeSiO_3$, 0.13) 1.83	0.60; 0.00; 0.60	2.10; 1.19; 3.29	1.20; 0.65; 1.85	0.40; 0.79; 1.19
Calcite	2.10	3.70	6.00	1.25	.....
Apatite	0.31	tr.	tr.	0.09	1.68
Dolomite	.....	.....	.....	.....	3.68
Water	0.74	0.79	2.12	1.45	2.35
Total	99.87	99.80	99.81	99.96	99.77

* $Ab_{95}An_5$. † $Ab_{98}An_2$. ‡ $Ab_{100}An_0$. § $Ab_{90}An_{10}$. ‖ $Ab_{89}An_{11}$.

I. Gray quartz-monzonite porphyry (462-*C*-7), Rock Creek; analyst: R. B. Ellstad.

II. Brown quartz-monzonite porphyry (462-*C*-40-*e*), Timberline Creek, Beartooth Plateau; analyst: R. B. Ellstad.

III. Gray quartz-monzonite porphyry (462-*C*-40-*d*), Timberline Creek, Beartooth Plateau; analyst: R. B. Ellstad.

IV. Quartz-keratophyre porphyry, east side Round Mountain, Nye quadrangle; analyst: T. Kameda.

V. Dacite porphyry, ridge northeast of Round Mountain, Nye quadrangle; analyst: T. Kameda.

with quartz. The orthoclase phenocrysts, $\frac{1}{4}$–$1\frac{1}{2}$ inches long, are fresh and euhedral. The mafics which were present as phenocrysts are completely altered to chlorite. Quartz, as clear fragments or partially resorbed doubly terminated hexagonal crystals, may be present as phenocrysts, but most of it is interstitial. The groundmass may or may not be altered and is generally composed of quartz and albite. The mode of this porphyry follows: quartz, 1.70 per cent; orthoclase, 14.45 per cent; albite, 25.84 per cent; biotite, 6.95 per cent; groundmass, 51.06 per cent. Figure 6 shows the polished surface of a hand specimen of gray porphyry.

The brown porphyry is generally badly altered and contains phenocrysts of clear glassy sanidine up to $\frac{1}{4}$ inch in diameter; zoned, altered albite; and some altered ferromagnesian mineral, probably pyroxene. No quartz is visible as phenocrysts. The groundmass may contain some areas of calcite but for the most part is composed of occult albite and quartz.

Both the red and brown porphyries are quartz monzonites. Chemical analyses and norms of these are given in Table 1.

These porphyry bodies of the Beartooth Plateau region have all the characteristics of intrusions formed under conditions of local tension rather than compression.

SUMMARY OF PETROGRAPHY AND PETROLOGY OF THE PORPHYRIES

The foregoing descriptions and the chemical analyses and norms given in Table 1 show that the porphyries do not represent widely divergent rock types. Texturally they are all similar in that the phenocrysts are embedded in a felsitic groundmass which is too fine grained to permit determination of its mineral constituents. The dominant phenocrysts are always plagioclase ranging in composition from $Ab_{90}An_{10}$ to $Ab_{70}An_{30}$, the high soda content being indicated in the chemical analyses and the normative plagioclase varying between $Ab_{100}An_{0}$ and $Ab_{89}An_{11}$. The preponderance of plagioclase over orthoclase is most conspicuous in all the porphyries and the chemical analyses show a low lime content. The presence of small quantities of ferromagnesian minerals also finds expression in the analyses, as iron and magnesia are present in relatively small amounts. Mineral-

ogically the chief difference between the porphyries is the presence or absence of quartz phenocrysts. Thus it would seem that all the porphyries just described are representative of a fairly normal intermediate to felsic magma.

AGE OF THE PORPHYRIES

East of the palisades of Line Creek the Wasatch (Eocene) beds of the adjacent Bighorn basin are composed of coarse conglomerate layers 10–15 feet thick which alternate with red and gray shales varying between 5 and 10 feet in thickness. Abundant porphyry pebbles, exact duplicates of the porphyries along the front and in the high plateau to the west, were found in the conglomerate layers. Besides the porphyries, pebbles of Paleozoic and pre-Cambrian rocks were also found. The proportions of these various pebbles are set forth in the following summary based on pebble counts in the Wasatch: pre-Cambrian (granite, gneiss, diabase), 50 per cent; porphyry, 20 per cent; Paleozoic (sandstone 15 per cent, limestone 2 per cent, shale 1 per cent), 18 per cent. These observations prove that most of the porphyries were intruded in pre-Wasatch time.

Despite the fact that the Paleozoic rocks along the Red Lodge front in the vicinity of Rock Creek have been folded and faulted, the porphyry maintains the same stratigraphic position and thickness throughout the length of the intrusive; the phenocrysts in the porphyry show a rather even distribution; and the primary structure of the porphyry, as indicated by the linear parallelism of the phenocrysts, would seem to indicate intrusion parallel to the bedding planes of the sediments. This, together with the displacement of the porphyry along the tear faults (Fig. 2), suggests intrusion as a horizontal sheet (previous to deformation of the Paleozoic rocks) in the first incompetent horizon (here the Gros Ventre shale) encountered by the magma rising through the pre-Cambrian and into the sediments. The sheetlike form, rather than lenticular or phacolithic, would also favor intrusion previous to, rather than during or after, deformation.

The relations of the porphyries in the Nye and McLeod areas indicate intrusion previous to, rather than during or subsequent to, the thrusting. Some laccoliths in the Nye area have arched up beds

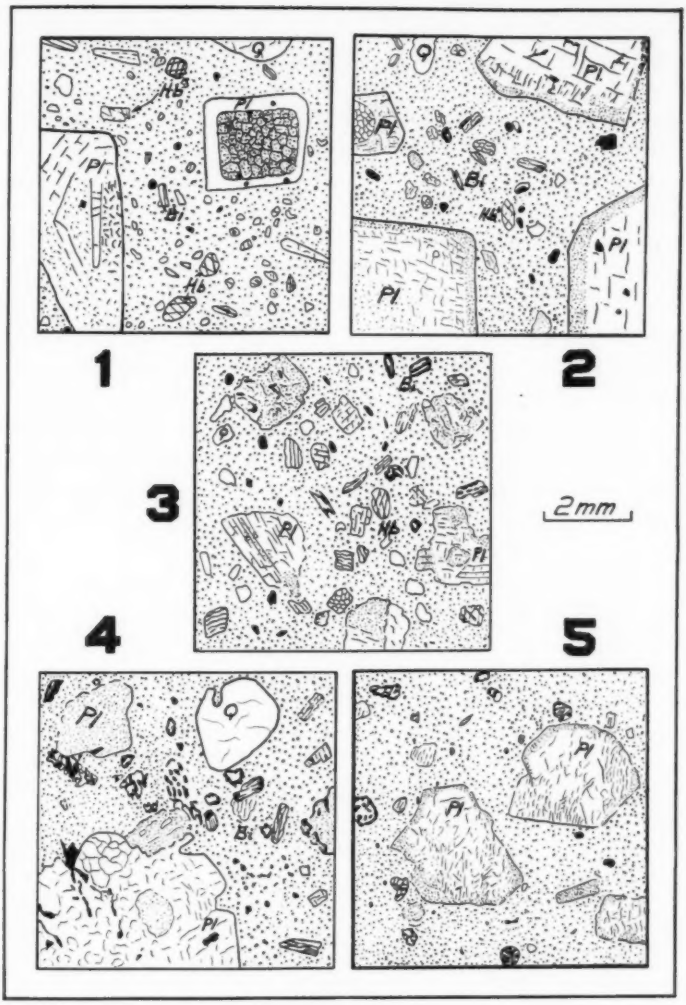

FIG. 7.—Drawings made from photomicrographs to show the characteristic texture of the porphyries. Phenocrysts are dominantly plagioclase which may or may not be altered. Clear crystals are quartz and the stippling represents the groundmass. *1*, gray quartz-monzonite porphyry, West Fork of Rock Creek; *2*, gray quartz-monzonite porphyry, Willow Creek; *3*, quartz-diorite porphyry, Elk Mountain; *4*, gray quartz-monzonite porphyry, East Boulder Plateau; *5*, brown quartz-monzonite porphyry, Beartooth Plateau.

as high in the section as Upper Cretaceous and therefore must be at least as young as those beds. The basal portion of the Livingston series of andesitic volcanic rocks contains a flora which Professor Dorf has identified as Upper Cretaceous and this series in places bevels and lies unconformably on the eroded flanks of the laccoliths. The dacite breccia near Enos Mountain lies immediately below the lower part of the Livingston series and above the Eagle sandstone (base of Montana group, Upper Cretaceous). Since the breccia is certainly the age equivalent of the dacite in the plug, the age of the porphyries of Enos Mountain is also nicely bracketed as Upper Cretaceous.

BROADER REGIONAL RELATIONS, TECTONICS, AND PORPHYRY INTRUSION

From the beginning of Paleozoic time up to the middle of the Cretaceous the Beartooth-Bighorn region was one of relative stability characterized by slow deposition of marine sediments though punctuated, particularly in the closing stages of this cycle, by various epeirogenic movements and oscillations from marine to terrestrial conditions. Faulting and folding during this whole length of time were almost entirely lacking. But during the Cretaceous, domal uplifts, warps, and movements along old zones of weakness in the pre-Cambrian basement began to take place. It has been suggested that elongate welts rose first as flexures and finally broke along their margins in steep faults. Recurrent compression caused pinching of the uplifted masses so that thrusting occurred particularly under the higher side of slightly tilted uplifted areas.[10] Some of this thrusting may have occurred near the end of the Cretaceous period but most of it was in the early Eocene (post–Fort Union but pre-Wasatch). In this paper it is referred to as the "Laramide thrust faulting" which is recognized as only one phase of the Laramide orogeny.

The porphyries definitely began to rise before the Laramide thrust faulting on the northern and eastern flanks of the Beartooth Mountains. Since the porphyries were emplaced before that thrusting

[10] W. T. Thom, Jr., C. W. Wilson, Jr., D. J. MacNeil, and D. L. Blackstone, Jr., "Results of Recent Studies of Certain Critical Structural Type, Space and Time Relationships in Yellowstone-Beartooth-Bighorn Region," *Proc. Geol. Soc. Amer. for 1933*, p. 58.

they were, therefore, independent of it; but they may well have been related to the bulging of the Beartooth range which preceded the horizontal thrusting on its border. Intrusions were, in most instances, controlled by the pre-existing doming and particularly by such domes as the Nye-Bowler lineament. The latter probably represents deep and old fractures in the pre-Cambrian basement, as suggested by Thom,[11] along which horizontal and vertical movements took place, facilitating the rise of magmas and resulting in the concentration of intrusives along them.

Igneous activity in the region as a whole did not stop with the Laramide thrusting, though for some reason porphyry intrusion ceased along the northern and eastern flanks of the Beartooth Mountains. The Livingston volcanic series must, however, have been extruded from vents in the vicinity of the northeastern front of the Beartooths and it apparently interfingers to the north with the sediments of the Lance. A few small lamprophyre dikes of post-thrust age are found near Red Lodge, Montana.

On the south side of the Beartooth block, porphyries were intruded after, rather than before, the thrusting. The tectonic conditions accompanying their formation were different and this is reflected in the shapes of the bodies themselves. J. T. Wilson[12] has shown that in the Mill Creek area the porphyries were intruded chiefly along steep faults which delimit the southern boundary of the Beartooth block.

In no case, either in the southern (younger) or northern (older) group of porphyry intrusions, is there evidence that they came up the thrust planes during the thrusting. As would be expected, the condition of compression along the thrusts at the time of thrusting seems to be unfavorable to advance of magma along these planes.

CHEMICAL COMPOSITION OF THE IGNEOUS ROCKS AND TIME SPACE RELATIONS

The porphyries of pre-Laramide thrust age, of the northern and eastern flank of the Beartooth Mountains, are representative of

[11] W. T. Thom, Jr., "Relations of Deep-Seated Faults to Surface Structural Features of Central Montana," *Amer. Assoc. Pet. Geol. Bull.* 7 (1923), pp. 1–13.

[12] "Geology of the Mill Creek-Stillwater Area, Montana," unpublished Ph.D. thesis, Princeton University (1936).

fairly normal intermediate to felsic magmas. They do, however, show a preponderance of soda-rich plagioclase over orthoclase even in the most felsic types. Thus quartz keratophyre occurs rather than the normal rhyolite of the calc-alkaline series. The absence of any mafic magmas should be noted. The Livingston volcanics, some of whose outcrops are in the general vicinity of the porphyries, are younger than the porphyries and are somewhat more mafic (pyroxene andesites).

Tertiary intrusives, younger than the Laramide thrusting, occur in a zone extending along the southwest side of the Beartooth and Absaroka Mountains. Some described areas along this zone from the northwest near the Yellowstone River southeast into the Absaroka Mountains include: Mill Creek, dacite, diorite;[13] Haystack Stock, granodiorite, quartz-bearing diorite, gabbro;[14] Cooke City, gabbro, monzonite, syenite, basalt;[15] Hurricane Mesa, basalt, gabbro, diorite, andesite;[16] Sunlight, andesite, diorite, basalt;[17] Deer Creek, quartz diorite.[18] These poryphries represent more divergent rock types than do the older porphyries and were derived from normal intermediate to mafic magmas. The volcanic rocks in these localities are predominantly mafic, corresponding to andesites and basalts, and in minor instances some of these extrusives and occasional small intrusive dikes or sills are abnormally high in alkalies. These high-alkali types make up the absarokite-shoshonite-banakite series of Iddings.[19]

In other areas, especially Yellowstone National Park, the effusive

[13] *Ibid.*

[14] William H. Emmons, "Geology of the Haystack Stock, Cowles, Park Co., Montana," *Jour. Geol.*, Vol. XVI (1908), pp. 193–229.

[15] T. S. Lovering, "The New World or Cooke City Mining District, Park Co., Montana," *U.S. Geol. Surv. Bull. 811* (1930), pp. 1–87.

[16] J. P. Iddings, "The Dissected Volcano of Crandall Basin, Wyoming," *U.S. Geol. Surv. Mono. 32*, Part II (1899), pp. 215–69.

[17] Willard H. Parsons, "Volcanic Centers of the Sunlight Area, Wyoming," *Geol. Soc. Amer. Prelim. Titles and Abstracts* (1936), pp. 36–37.

[18] John T. Rouse, "The Structure, Inclusions and Alteration of the Deer Creek Intrusive, Wyoming," *Amer. Jour. Sci.*, Vol. XXVI (1933), pp. 139–46.

[19] J. P. Iddings, "Absarokite-Shoshonite-Banakite Series," *Jour. Geol.*, Vol. III (1895), pp. 935–59; John T. Rouse, "The Volcanic Rocks of the Valley Area, Park Co., Wyoming," *Trans. Amer. Geophys. Union, 16th Ann. Meeting* (1935), pp. 274–84.

rocks are characterized by great rhyolite flows and smaller basalt flows. These represent the last stages of igneous activity which

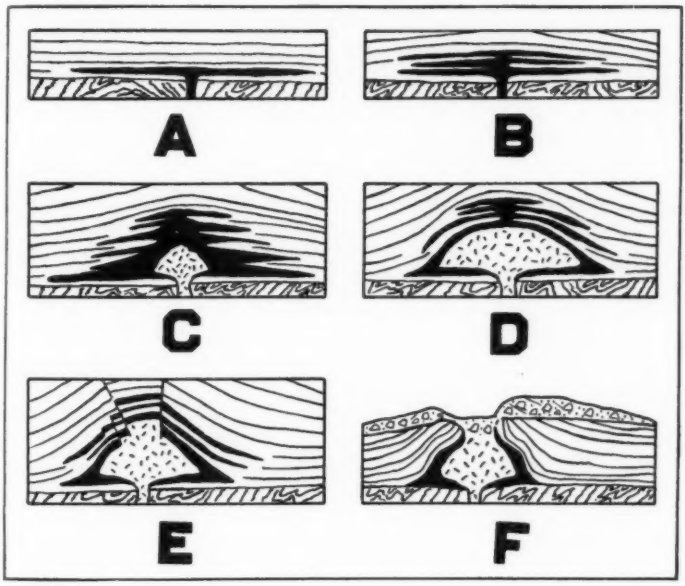

FIG. 8.—Diagrammatic representation of the shapes of the various pre-Laramide thrust porphyries showing lines of evolution from sills to laccolithic bodies and volcanic vents. *A* and *B*, Sills-Red Lodge front, Beartooth Plateau; *C* and *D*, Christmas-tree laccoliths—Nye, Lodgepole Creek; *E*, bysmalith—Nye; *F*, volcanic vent—Enos Mountain.

started late in the Tertiary and has continued almost to the present time.

CONCLUDING STATEMENT

The present distribution of the porphyry intrusions around the margins of the Beartooth Mountains suggested, before this study was made, that they were related perhaps to the thrust faults which delimit the mountain block. As shown in the foregoing discussion, this idea has been discarded as untenable. The close areal relations of porphyry intrusions to the thrust faults is perhaps chiefly dependent on the fact that the most favorable loci of intrusion are

along the Cambrian–pre-Cambrian contact or in incompetent shales within the Cambrian, and that this zone is now most extensively exposed in the vicinity of the thrust faults near the mountain borders. Had erosion gone less deep, porphyries would probably be found above the crystallines now surfacing the lofty Beartooth Plateau, and had it gone deeper they would be found presumably under the sediments of the present adjoining basins.

SMALL GRANODIORITIC BLOCKS FORMED BY ADDITIVE METAMORPHISM[1]

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ABSTRACT

Small irregular granodioritic blocks entirely surrounded by quartz-biotite hornfels are well exposed in a mine tunnel in the Cornucopia district of northeastern Oregon. These occur in certain zones in which the hornfels exhibits marked cataclastic structures, but none of the granodioritic blocks were sheared even though they show sharp outlines. Instead some of them have gradational contacts with the hornfels and contain inclusions of that rock. Apophyses extend into the hornfels from the margins of some of these granodioritic blocks.

The relationship cannot be adequately explained by the usual interpretations of a cataclastic breccia, or of magmatic injection, or of the faulting of small granodiorite intrusions. Field relations and the study of many thin sections point to the conclusion that these granodioritic blocks are the result of recrystallization replacement of highly crushed portions of the hornfels, and that their shape is, in part, controlled by the angularity of the crushed hornfels.

INTRODUCTION

In the late summer of 1935, during the driving of a long, low-level adit, called the Coulter Tunnel, at Cornucopia, Oregon, small angular blocks of granodioritic rock in a matrix of hornfels were encountered at a distance of some 4,300 feet from the portal and at a depth of some 1,200 feet below the surface. This occurrence was unusual enough to attract the attention of the men in charge of the tunnel operations, who were much surprised at being able to mine out blocks of "granite" surrounded by "greenstone." These blocks have a striking resemblance to xenoliths but are just the reverse in that they consist of blocks of fresh, unmetamorphosed granodioritic material in a schistose hornfels (Fig. 1). The miners called the occurrence "pinto rock."

The Cornucopia mining district is situated on the southeastern flanks of the rugged Wallowa Mountains, one of the higher ranges of the Blue Mountain group of northeastern Oregon. Although de-

[1] Presented under the title of "The Mode of Origin of Small Granodioritic Blocks" at the April, 1936, meeting of the Cordilleran Section of the Geological Society of America, Pasadena, Calif.

tailed geologic maps of this area are lacking, several general papers on this region have been published.[2] In addition, factual data concerning the size and shape of rock masses gained from excellent surface exposures are supplemented by those obtained from extensive mine workings.

FIG. 1.—Flashlight photograph taken in the "Coulter" tunnel showing the occurrence of granodioritic blocks in crushed hornfels.

The salient features of the geologic history of this region might be listed as follows: first, pre-Tertiary sediments and volcanics now metamorphosed to schists, hornfels, and greenstones; second, the emplacement of granodioritic and associated rocks such as porphyries,

[2] W. Lindgren, "Gold Belt of the Blue Mountains of Oregon," *U.S. Geol. Surv., 22d Ann. Rept.*, Part II (1900–1901), pp. 561–776; James Gilluly, S. C. Reed, and C. E. Park, Jr., "Some Mining Districts of Eastern Oregon," *U.S. Geol. Surv., Bull. 846A* (1933).

granophyres, aplites, pegmatites, and quartz veins; third, development of a mature erosion surface with maximum relief of 3,000 feet; fourth, profound fracturing forming conduits for extensive flows of basaltic lavas which completely flooded the Middle-Miocene land surface; fifth, block faulting probably in Pliocene time, which according to Livingstone,[3] has had a marked control on present-day topography; sixth, vigorous alpine glaciation during the Pleistocene. Deep U-shaped valleys afford cross sections of 2,000 to 3,000 feet.

In the new low-level Coulter Tunnel, hornfels of pre-Tertiary age is the prevailing rock. This hornfels varies in color from dark green to black, but where exposed on the surface the oxidation of disseminated pyrite gives rise to various shades of reddish brown. In general the hornfels is compact, in part slightly schistose, and commonly exhibits a blocky appearance caused by numerous intersecting joints. In the tunnel it is traversed by dikelike masses of aplite and stocklike masses of granodioritic composition. There are also steeply dipping zones which show marked effects of crushing and very evident cataclastic structures. Such zones may be 50 feet or more in width. In them the slightly schistose hornfels has been sheared and fractured in such a way that angular fragments are clearly visible. That many of the fragments have suffered rotation is evidenced by the abrupt changes in the direction of the schistosity of adjacent fragments.

Locally in driving the tunnel through one of these zones the finely comminuted material was found to be unconsolidated, but in most of the crushed hornfels the fragments are tightly compacted. In these crushed zones there are some smooth slickenside surfaces which are characterized by the development of chloritic material but which have no uniform orientation. The compacted rock contains a few small vuglike cavities which are partially filled with minute turbid crystals of calcite. This mineral also occurs as thin films between the fragments of the hornfels and is usually associated with pyrite. The crushed hornfels presents a very coarse, hackly appearance, and broken surfaces show many angular pits from which fragments have been detached (Fig. 2).

[3] D. C. Livingstone, "Certain Topographic Features of Northeastern Oregon and Their Relation to Faulting," *Jour. Geol.*, Vol. XXXVI (1928), pp. 694–709.

It is probable that these crushed zones of cataclastic metamorphism have a considerable vertical extent. Similar zones have been observed in the upper workings and on the surface some 1,200 feet above the adit. That they may continue down from these workings

FIG. 2.—Crushed hornfels showing pits between fragments. In the upper portion of this specimen there is a "porphyritic" block containing inclusions of the hornfels. Part of the contacts are gradational. Approximately two-thirds natural size.

is indicated by the large amount of surface water coming through them in the lower tunnel.

PETROGRAPHY OF THE HORNFELS

Petrographic study of many specimens from the tunnel and the surface shows that in microtextures and structures the hornfels varies from the finer-grained varieties where the interlocking crystals are so small as to be barely distinguishable to the coarser varieties where the microtexture approaches that of a granulite. Most sections exhibit a xenoblastic intergrowth of unstriated plagioclase, quartz, biotite, and hornblende. Biotite and hornblende commonly show alinement and either mineral may predominate. In some sections, however, a decussate structure is present, and a few sections appear

to be typical *Garbenschiefer* with plumed porphyroblasts of anthophyllite. Other minerals present in various sections of the hornfels are garnet, diopside, epidote, titanite, calcite, chlorite, magnetite, and pyrite. Garnet, diopside, and epidote occur in irregular patches and lenticular veinlets which are probably similar in origin to the quartz-diopside-garnet veinlets common to this vicinity.[4]

Porphyroblastic textures are very common. Some of the hornblendic facies contain many feldspar porphyroblasts 2 or 3 mm. in size so that megascopically the hornfels has the appearance of a melanocratic porphyritic rock, but under the microscope their porphyroblastic nature is obvious. Sieve structures, helizitic structures, and all gradations in the development of the various porphyroblasts are common. These features are especially well developed in many of the plagioclases. Some of the porphyroblasts exhibit a turbidity in all stages of development, while others are clear even in the initial stages. The included material which produces the turbidity may have been some of the original finest constituents of the rock, or it may have been kaolinitic material caused by hydrothermal alteration just prior to the formation of the porphyroblasts.

In the earliest stage of growth, the feldspar porphyroblasts are decidedly anhedral, some even amoeba-like in form, surrounding and including all other constituents; then as crystallization proceeds they become subhedral, and in these stages kaolinitic material may be evenly distributed throughout the crystals. Then the turbid kaolinitic material appears to be segregated usually in the central portions of the crystals while in the last stages it is irregularly distributed or segregated along twinning planes. The segregation of kaolinitic material may be analogous to the movement of inclusions by crystallization pressure.[5] Kaolinitic material thus associated with porphyroblasts bears some resemblance to the product of subsequent hydrothermal alteration, but may be differentiated from it because the effect of the infiltration of subsequent alteration material either along cleavage cracks of originally clear crystals or in traversing veinlets is usually easily recognized by its distribution and by its mineral

[4] G. E. Goodspeed and Howard Coombs, "Quartz-Diopside Garnet Veinlets," *Amer. Mineral.*, Vol. XVII (1932), pp. 554–61.

[5] A. Harker, *Metamorphism* (London: Methuen & Co., Ltd., 1932).

association. Material produced by the usual processes of hydrothermal alteration is commonly associated with clear unaltered minerals.

Porphyroblasts of garnet are usually very irregular in outline and give the suggestion that they may have replaced chlorite. Some of the hornblende porphyroblasts contain abundant inclusions in helizitic arrangement and have formed at nearly right angles to the general alinement of the earlier minerals. Xenoblastic aggregates of

TABLE 1

CHEMICAL ANALYSES OF HORNFELS AND OF A GRANITOID BLOCK*

Constituent	Hornfels *C*106–35	Granitoid Rock *C*112–35
SiO_2	50.12	70.03
Al_2O_3	18.69	16.27
Fe_2O_3	5.47	0.62
FeO	6.48	1.60
MgO	4.34	0.82
CaO	5.90	4.16
Na_2O	4.17	4.56
K_2O	1.90	0.83
H_2O—105° C.	0.65	0.25
H_2O—105° C.	0.90	0.70
P_2O_5	0.10	0.11
TiO_2	0.64	0.24
MnO	0.18	Nil
Total	99.54	100.19

* Analyses by Frank Herdsman of Glasgow through funds provided by the Research Committee, University of Washington.

epidote are present in the less siliceous facies. Quartz is also xenoblastic in the hornfels exposed in the Coulter Tunnel although less than a mile to the north idioblastic quartz is found in a siliceous hornfels.[6]

Many thin sections from the crushed portion of the hornfels show on a small scale the effects of crushing which are so readily observable in the field or in the hand specimen. In certain facies in which granulation and flaser structures are present the change in the direction of the schistosity testifies to the rotation of the small angular

[6] G. E. Goodspeed, "Development of Quartz Porphyroblasts in a Siliceous Hornfels," *Amer. Mineral.*, Vol. XXII, No. 2 (Feb., 1937).

blocks of hornfels. Interlocking crystals bordering many of the fragments weld them tightly together although in the matrix there is usually an increase in grain size and an increase in quartz which is commonly accompanied by twisted and contorted flakes of biotite. In many of the broken blocks of hornfels there is a very noticeable development of small feldspar porphyroblasts, and some fragments

FIG. 3.—Angular granodioritic blocks. Close inspection will reveal small projections of the hornfels into the granodiorite and small extensions of the granodiorite into the hornfels. What appear like fragments are in reality porphyroblasts or aggregates of porphyroblasts. Natural size.

adjacent to one another have striking changes in mineral composition. In a single thin section a biotite-rich fragment may be adjacent to one in which most of the biotite has been replaced by hornblende. Hornblende in some of these sections may show an increase in grain size, and crystals over 0.5 mm. are common. In other sections there are zones or areas rich in magnetite distributed as individual grains or in rounded patches liberally sprinkled with very finely divided

particles. A chemical analysis of the sheared hornfels (*C*106–35) shows a high alumina content, abundant iron, lime, and magnesia, over twice as much soda as potash, and a silica content of about 50 per cent (Table 1).

GRANITOID BLOCKS

The crushed zones of dark, fine-grained, schistose hornfels would not be especially noticeable in the Coulter Tunnel were it not for the fact that they are studded with angular blocks and veinlets of light-

FIG. 4.—Hackly hornfels with pits. Granodioritic material in veinlike occurrences and surrounding fragments of the hornfels. Approximately two-thirds natural size.

colored granitoid rock, the gleaming whiteness of which is in striking contrast to the dark matrix (Fig. 3). Some of the veinlets fill in around fragments of the hornfels (Fig. 4). The blocks have no particular alinement or arrangement but are distributed haphazardly, some in close groups, others more isolated. They range in size from a fraction of a centimeter to over half a meter in diameter, with an average size of about 5 or 10 cm. (Fig. 5). Most of the blocks are leucocratic, but some which are much darker in color are not readily distinguished from the hornfels unless, as is usually the case in the tunnel, the rock is wet (Figs. 6 and 14).

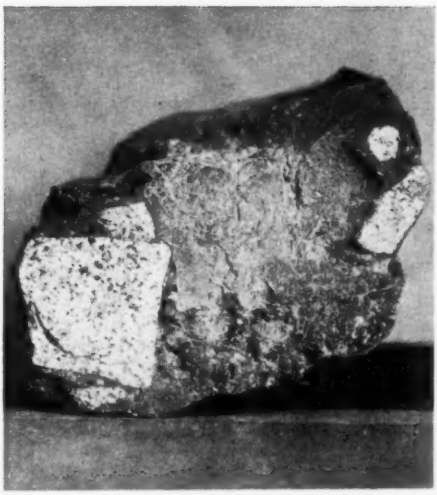

FIG. 5.—The largest block in this specimen shows a rectangular section of approximately 2½ inches. There are numerous porphyroblasts in the hornfels, and the smaller block immediately above the larger one shows a gradational contact.

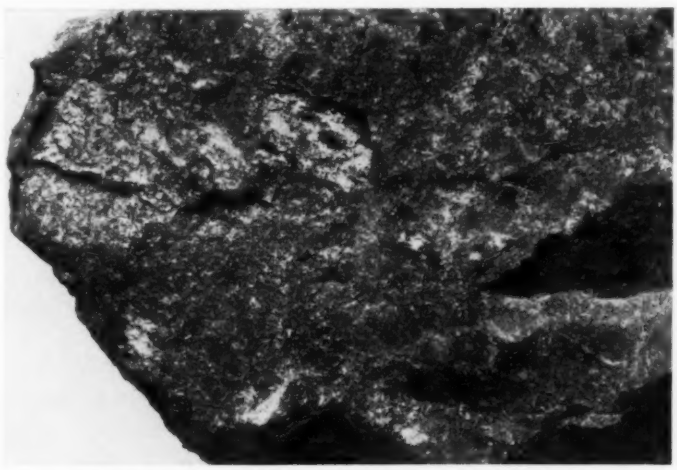

FIG. 6.—Hackly hornfels with a "porphyritic" block showing ill-defined contacts. An angular depression is but partially filled with quartz, feldspar, and fragments of hornfels. Approximately three-fourths natural size.

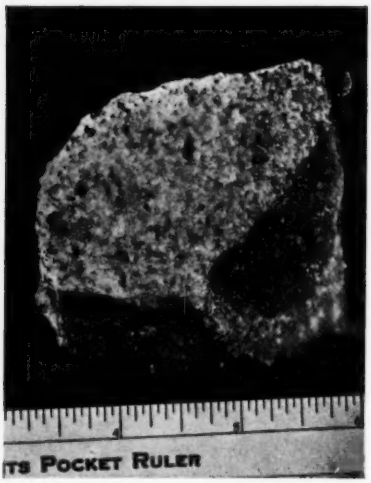

FIG. 7.—Portion of a block (Specimen *C*112–35) which is a quartz-diorite in composition. Note the irregular extension of the granitoid rock into the hornfels.

FIG. 8.—Small irregular-shaped block with veinlike extensions. Natural size

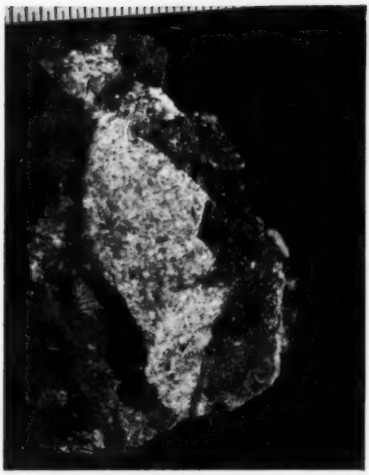

FIG. 9.—Small granodioritic block with a sliver of hornfels and thin extensions of granodioritic material protruding into the hornfels. There are no apparent lines of fracture within this block. Two-thirds natural size.

FIG. 10.—Granodioritic block which in part exhibits gradational contacts. Natural size.

FIG. 11.—Granodioritic block showing graduational contacts, inclusions of hornfels, and veinlike extensions of granodioritic material protruding into the hornfels. Natural size.

FIG. 12.—Relicts of recrystallized hornfels septa in a block which was probably formed by the coalescing of several smaller blocks.

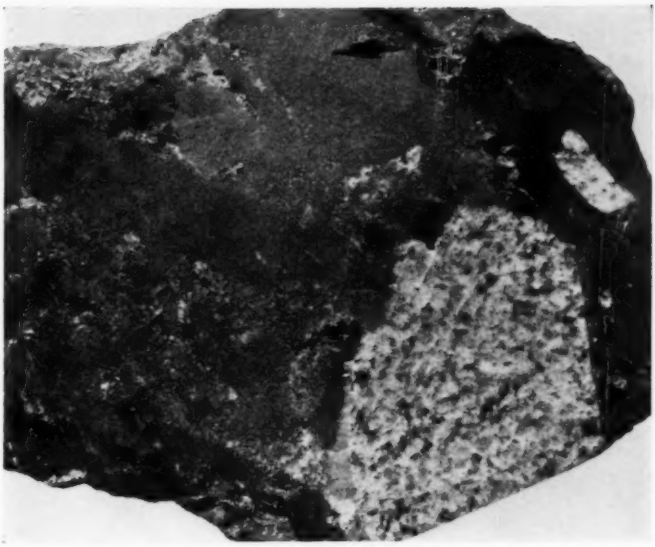

FIG. 13.—Polished section of a granodiorite block in crushed hornfels. Incipient aggregates of granodioritic material between the fragments of hornfels. Natural size.

FIG. 14.—Porphyritic block barely discernible from the surrounding hornfels. One-half natural size.

Although sharp borders are most noticeable, some of the blocks have irregular outlines with protuberances into the hornfels and also contain slivers of hornfels in granitoid rock, while others show very evident gradational contacts with the hornfels (Figs. 7, 8, 9, 10, and 11). Several of the larger blocks contain narrow bands of hornfels

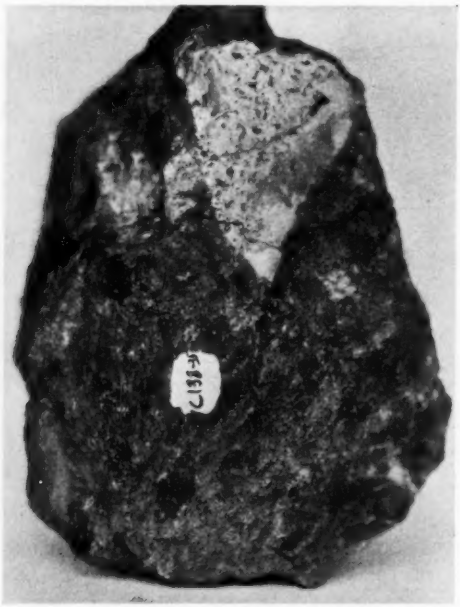

FIG. 15.—Pegmatitic type of block with a vuglike cavity. Three-fourths natural size.

in a pattern which suggests the growing-together of several smaller blocks (Fig. 12). Many of the smaller ones are similar in outline to the angular depressions found on the hackly surface of the severely crushed portion of the hornfels (Fig. 13). However, it is clearly evident that the granitoid blocks have not suffered the cataclastic metamorphism which is so apparent in the hornfels. The blocks have a decidedly fresh, unaltered appearance and in general a medium-to-

coarse granitoid texture. Plagioclase and quartz are usually dominant minerals with biotite as the chief accessory mineral and with magnetite and pyrite as very minor constituents. Some of the darker-colored blocks have porphyritic textures while a few of the more leucocratic ones show pegmatitic tendencies (Fig. 14). A few blocks contain small irregular vuglike cavities (Fig. 15). Some of them have

FIG. 16.—Granodioritic blocks occupying interstitial positions with respect to the fragments of hornfels. One of the blocks has a veinlike apophysis of granodioritic material. One-half natural size.

narrow veinlets of granodioritic material as apophyses, and some of the blocks contain inclusions of the hornfels (Figs. 11 and 16).

Petrographic studies of many thin sections of the granitoid blocks show a hypidiomorphic granular texture with subhedral plagioclase (about 1 mm.), anhedral quartz (about 0.5 mm.), and shreds of brown and green biotite which is in part altered to chlorite and is associated with magnetite and titanite as minor constituents. In the porphyritic blocks the larger feldspars are set in a finer-grained aggregate of quartz, feldspar, and biotite. In general the plagioclase

is andesine with about 35 per cent An and usually exhibits oscillatory zoning and various kinds of twinning. The feldspars have sutured boundaries, and some of the individual crystals have the appearance of a coalesced aggregate of several smaller feldspar crystals.

Several of the plagioclase crystals have included turbid patches of kaolinitic material which do not appear similar to the products of subsequent alteration. Some of the feldspars include small grains of quartz, flakes of biotite, and occasional grains of titanite, apatite, and white garnet. A few of the plagioclase crystals appear to be replaced by orthoclase, and many of them are surrounded by a clear rim of much more sodic composition than the central portion. There are a few micropegmatitic intergrowths which suggest the replacement of feldspar by quartz, which is usually interstitial. The quartz contains some vacuole inclusions, commonly concentrated near the borders of a feldspar which shows a replacement contact. Perhaps these vacuole inclusions may bear a genetic relation to the mechanism of replacement. Similar features were noticed in a petrographic study of the gold-quartz veins of Cornucopia.[7]

Carbonate is present in some of the sections and occurs in several ways. A few of the plagioclases contain calcite included in the central portion of zoned crystals. It is doubtful if this carbonate is a subsequent alteration product not only on account of its position but because the feldspar is clear and unaltered. Calcite also occurs interstitially in a few sections in a position similar to quartz and makes sharp boundaries with unaltered feldspars. In one section there is some calcite included in the plagioclase feldspar forming a peculiar micrographic intergrowth which suggests a replacement of the feldspar by the carbonate. There are also a few veinlets of carbonate which are clearly subsequent to the formation of the other minerals. No evidence of dynamic metamorphism is present.

Although these blocks vary in texture and composition, they appear in the main to be sodic granodiorites and quartz-diorites. Plagioclase is in excess of orthoclase, and biotite is the dominant mafic mineral. The quartz-diorites have an average approximate mineral composition of about 65 per cent plagioclase, 25 per cent quartz, and 10 per cent of other minerals. A chemical analysis (*C*112–35) of a

[7] G. E. Goodspeed, "Microstructures and Metallization of the Gold-Quartz Veins of Cornucopia, Oregon," *Econ. Geol.*, Vol. XXXI, No. 4 (June-July, 1936), pp. 398, 416.

specimen of a medium-grained, unaltered granitoid block shows a composition more typical of quartz-diorite than a normal granodiorite since there is a much greater amount of soda than potash. The small H_2O content testifies to the freshness of the rock.

CONTACT RELATIONS

An attempt was made to secure many data regarding the contact relations of these blocks since such evidence is most pertinent to any

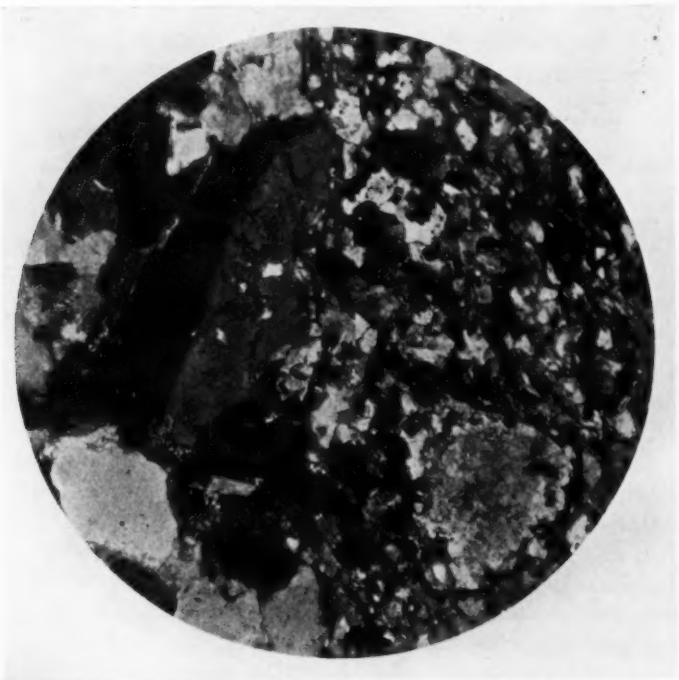

FIG. 17.—Photomicrograph of a contact between a granodioritic block and hornfels. There is a turbid incipient feldspar porphyroblast in the hornfels. A plagioclase crystal about 1 mm. long directly on the contact shows crenulated borders with the hornfels and a partial development of sieve structure.

interpretation regarding their mode of origin. There is not the slightest indication of any chilled selvages or finer-grained borders. Plagioclase crystals of normal size in the granitoid blocks occur along the

contact of the schistose hornfels, and these crystals commonly extend into the fine-grained hornfels and include some of the finer-grained constituents, thus producing a sieve structure (Fig. 17). Some sections show a marked development of porphyroblasts of plagioclase and quartz in the hornfels immediately adjacent to the contact. There are complete gradations from a porphyroblastic hornfels to what would be called a quartz-diorite porphyry. Some sections show the inclusion of remnants of the hornfels in the granitoid portion. Some of these remnants are angular or irregular fragments of hornfels; others exhibit delicate lacelike gradations into the granitoid rock. In many sections the contact of the hornfels is marked by an irregular increase in the distribution of biotite, chlorite, and magnetite. The schistosity of the hornfels with its abrupt changes may be at any angle to the contact of the granitoid block.

ORIGIN

One of the hypotheses which might be suggested for the mode of origin of these granitoid blocks is that they were originally of sedimentary origin, either a conglomerate or a clastic breccia. The distribution of the blocks, the lack of cataclastic effects in the blocks, the absence of deformation, the protuberances of the granitoid rock into the hornfels, and the inclusions of hornfels in the blocks are entirely against such an interpretation. The sharp angularity of most of the blocks is difficult to explain by sedimentary origin, but the transition of some of the blocks into veinlike forms makes this hypothesis untenable.

Another interpretation that would account for the angularity of the blocks is that they represent broken fragments of pre-existing masses of granitic rocks. According to this explanation original lit-par-lit or dikelike injections were later broken and displaced during the same period of metamorphism which produced the hornfels. Notwithstanding the simplicity and apparent plausibility of this interpretation, there are several features of these granodioritic blocks that are not explained by it. Small aggregates of quartz and feldspar or single crystals of feldspar which look like detached fragments exhibit in thin section cogent evidence of crystalloblastic origin (Fig. 18). Sieve structures and irregular sutured boundaries are common.

Groups of several crystals may enclose a portion of hornfels. Some of the blocks show distinctly gradational contacts and even relatively sharp contacts may exhibit, in thin sections, sharp knife-edge extensions of plagioclase into the hornfels (Fig. 19). Fragments of the hornfels within the blocks and the prolongation of the granodioritic material into the hornfels are against this interpretation.

It might be suggested that a magma was injected into the crushed hornfels along imperceptible joints and cracks. On this assumption

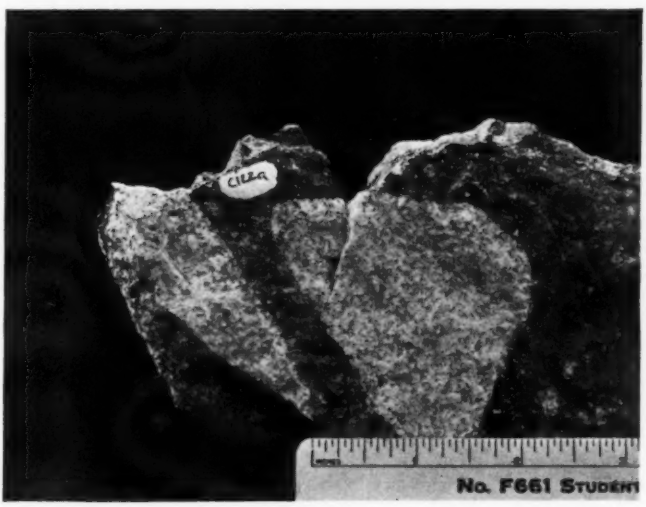

FIG. 18.—The septum between these blocks and the adjacent hornfels contains numerous porphyroblasts.

some evidence of flowage and alinement of the mineral constituents with reference to the periphery of the blocks would be expected, but such features are entirely absent. Even thin delicate fragments of hornfels within the blocks show no deformation. Most of the blocks are typically granitoid in texture and some of them contain irregular dark streaks rich in mafic minerals. However, it is believed that these streaks represent the remnants of formerly existing septa of hornfels between originally smaller granitoid blocks since such septa in various stages of recrystallization may be found. The association

of such minerals as pyrite, calcite, and garnet is somewhat incongruous to a magmatic interpretation. Another point against this interpretation is the presence in some of the blocks of small irregular cavities which are neither evenly distributed as in a diktytaxitic texture[8] nor lined with crystals as in miarolitic cavities. The major ob-

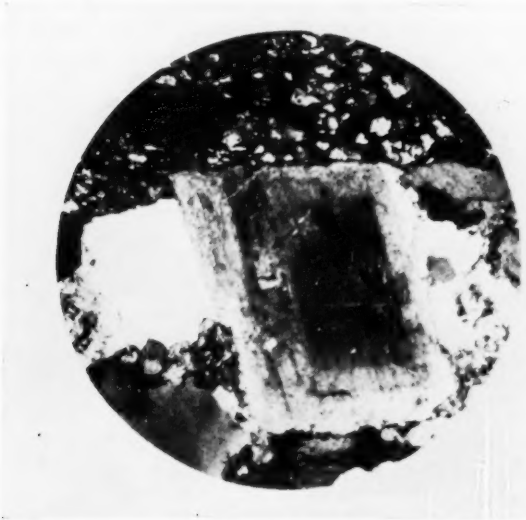

FIG. 19.—Photomicrograph of a contact of a granodioritic block and hornfels. There is some inclusion of hornfelsic material, and a plagioclase crystal exhibits a sharp, thin-edged projection into the hornfels. This crystal is 1.3×1 mm. in size.

jection to the magmatic interpretation, however, is found in the contact relations of these blocks to the hornfels.

Plagioclase crystals which extend from the granitoid blocks into the hornfels and there include the finer constituents in characteristic porphyroblastic manner indicate crystalloblastic origin for this mineral, which in turn suggests a similar mode of formation for all minerals in these blocks. There is no suggestion that they were in part formed from a true melt. The feldspars in the blocks have features

[8] R. E. Fuller, "The Geomorphology and Volcanic Sequence of Steens Mountain in Southeastern Oregon," *Univ. Wash. Publ. in Geol.*, Vol. III, No. 1 (November, 1930), pp. 1–130.

similar to those more fully developed porphyroblasts in the hornfels. As might be expected from this interpretation, blocks may be found in various stages of development. The more mafic blocks merely included a larger amount of hornfelsic material, and exhibit plagioclase porphyroblasts in an incipient stage of development.

In order to explain all features of these blocks it seems necessary to offer a more complicated hypothesis. If thermal emanations or hydrothermal solutions penetrated the crushed hornfels, it might be expected that they would replace most easily those portions where crushing had been severe. Here there would have been a greater permeability and a greater area of material to be acted upon, while the slickenside surfaces, perhaps less permeable, served as boundaries. Hence the angular form of the block may in part represent the feldspathization of angular crushed portions.

According to this hypothesis, the porphyritic blocks probably represent an initial stage of recrystallization, and the phenocrysts are really porphyroblasts. As recrystallization replacement proceeded, the divisions between some blocks became included in the larger block. The irregular bands of higher concentration of mafic constituents thus represent the remnants of septa. Many of the larger blocks may have resulted from the encroachment of smaller blocks on the dividing septa of hornfels so that several adjacent blocks with gradual growth merged into a larger one.

A comparison of the chemical analysis of the hornfels and the analysis of a granitoid block shows a decided increase in silica in the igneous-appearing rock and a decrease in magnesia, iron, and potash. This would indicate that as the crushed hornfels was recrystallized there was a transfer of certain constituents such as iron, magnesia, and potash from the areas of recrystallization, i.e., the blocks, into the surrounding hornfels. This would account for the increase in biotite, chlorite, and magnetite.

Among others, E. S. Bastin,[9] G. H. Anderson,[10] and James Gil-

[9] "Aplites of Hydrothermal Origin Associated with Cobalt-Silver Ores," *Econ. Geol.*, Vol. XXX, No. 7 (November, 1935), pp. 715–35.

[10] "Granitization, Albitization and Related Phenomena in the Northern Inyo Range of California-Nevada," *Geol. Soc. Amer. Bull.*, Vol. 48 (1937), pp. 1–74.

luly,[11] have noted the efficacy of hydrothermal replacement in the formation of rocks the origin of which is commonly assigned to the direct crystallization from molten magmas.

In a recent publication dealing with the transfusion of quartzite, D. L. Reynolds has pointed out that sharp contacts may represent diffusion limits and that the facts that a rock exhibits an igneous appearance both in the hand specimen and in thin section, and is of igneous composition, do not constitute evidence that it has crystallized from a magma.[12]

The term "recrystallization replacement" is here suggested to designate processes of hydrothermal metamorphism which may produce a fresh, unaltered granitoid rock in contrast with the usual product of subsequent hydrothermal alteration. Although the results of such processes may be more commonly expressed in such forms as replacement dikes and replacement breccias,[13] this occurrence of granodioritic blocks is of interest in displaying the apparent compatibility with respect to origin of either sharp or gradational contacts.

It is believed that the field and petrographic evidence justifies the interpretation that these granodioritic blocks were formed by recrystallization replacement as a process of additive metamorphism. Further studies in the Wallowa Mountains may give additional evidence as to whether or not this process of recrystallization replacement can be applied to the larger masses of granodioritic rock.

[11] "Replacement Origin of the Albite Granite near Sparta, Oregon," *U.S. Geol. Surv., Prof. Paper 175-C* (1933), pp. 65–81.

[12] Doris L. Reynolds, "Demonstration in Petrogenesis from Kiloran Bay, Colonsay. I. The Transfusion of Quartzite," *Mineral. Mag.*, Vol. XXIV, No. 105 (December, 1936), pp. 367–407.

[13] G. E. Goodspeed and Howard A. Coombs, "Replacement Breccias of the Lower Keechelus," *American Journal of Science*, Vol. XXXIV, July, 1937.

A QUANTITATIVE PETROGRAPHIC STUDY OF THE BLACK MOUNTAIN LEUCOGRANODIORITE AT WEST DUMMERSTON, VERMONT

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INTRODUCTION

The use of granite as a building stone has stimulated investigations of the many occurrences of this material in New Hampshire and Vermont, and the occurrence of "granite" at Black Mountain has been frequently mentioned and briefly described in several reports;[1] however, no detailed quantitative study of the minerals of this rock has been made.[2] It was, therefore, thought desirable to make such a study of the Black Mountain intrusive which is being used at present for structural purposes.

LOCATION AND DESCRIPTION OF THE AREA

The area is located in the southeastern part of Vermont in the town of Dummerston, about $5\frac{1}{2}$ miles northwest of Brattleboro (Fig. 1). A low-to-medium relief prevails in the region where the physiographic features trend in a general north-south direction in conformity with the structural features. The leucogranodiorite occurs on both sides of the West River at the village of West Dummerston where the river angles across the structure to join the Connecticut at Brattleboro. The northern and larger part of the intrusion is known as Black Mountain. It rises in the form of a crescent-shaped ridge to an elevation of 1,269 feet.

[1] T. Nelson Dale, "The Granites of Vermont," *U.S. Geol. Surv. Bull. 404* (1909), pp. 105–9. J. E. Maynard, "The Petrographic Re-examination of Quartz-bearing Plutonites from Vermont," *Jour. Geol.*, Vol. XLII, No. 2 (1934), pp. 148–50. Edward Hitchcock and others, *Report on the Geology of Vermont* (Claremont, N.H., 1861), Vol. I, p. 742.

[2] J. E. Maynard has given the mineral analyses of this intrusion but does not give the number of slides examined.

GENERAL GEOLOGY

The country rock into which the leucogranodiorite is intruded consists of a series of interbedded, somewhat gneissoid schists which

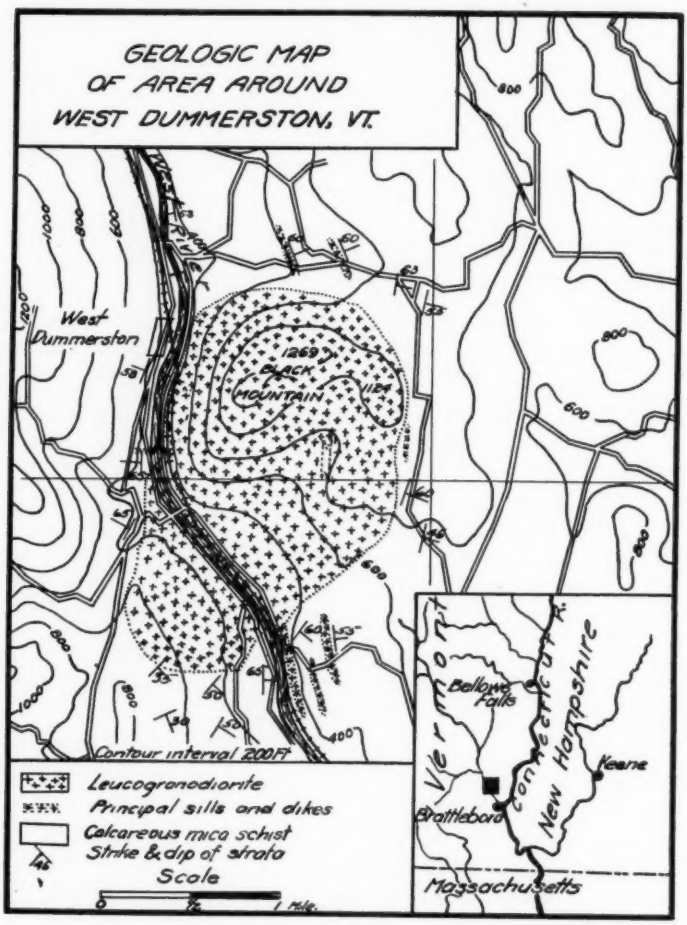

FIG. 1.—Geologic map of Black Mountain and adjoining area

vary greatly in texture and composition. Among these are strata of impure limestone, impure quartzite, sandstone, and mica schist.

Actinolite zoisite, and garnet are often important minerals in the various strata. Quartz veins, believed to be of metamorphic origin, are usually parallel to the beds, but in places they cut sharply across them. The United States Geological Survey's *Geologic Map of the United States*, published in 1932, describes the rocks of the area as phyllites, of unknown age, and places them below the Cambrian. The formations to the east and northeast in Vermont are assigned to a still earlier period. Billings[3] and his co-workers have shown that the latter are Silurian or early Devonian in age. The sediments in the vicinity of West Dummerston are probably early Devonian.

The strikes and dips of the schists about the intrusive are shown on the geologic map, from which it is seen that the general trend of the strata has not been greatly modified by the intrusion of the igneous material.

Along the east and west sides of the igneous mass, dikes and a few sills are common in the sediments over a distance of 100–200 feet; most of these are less than 2 feet across, but some are considerably larger. Contact metamorphic effects are not noticeable although there probably has been some interchange of material, for the contacts between dikes and sills, which are contemporaneous, and the country rock are not all of knife sharpness and small lit-par-lit stringers of igneous material are not uncommon. Some of the dikes that cut across the bedding are up to 20 feet in thickness. To the north and south of the intrusion the sediments are much obscured by drift, but where exposed they do not show a noticeable contact metamorphic effect. It is believed, therefore, that the sediments were metamorphosed before the time of intrusion and that the presence of the magma did not further change the composition of the surrounding rocks to any appreciable extent.

A short distance to the south of the crescent-shaped ridge of Black Mountain a large xenolith of sediments remains in its original north-south orientation. This xenolith shows a somewhat higher degree of lit-par-lit injection than the surrounding country rock.

THE INTRUSION AND ITS APOPHYSES

The plan (Fig. 1) and visible form of the intrusive indicate that it is a stock; and the field relations show it to be injected along the

[3] Marland Billings, *Geology of the Littleton and Moosilauke Quadrangles, New Hampshire* (Concord, New Hampshire, State Planning and Development Commission, 1935).

anticline of a fold. The area exposed is approximately 2 miles long by $1\frac{1}{2}$ miles wide, the longer dimension extending north and south. Several prominent sills occur to the north and south of the main mass. The largest one is 40 feet thick and 1,000 feet long. They stand out as prominent ledges through the glacial till.

The structure in the northern part of the main mass shows a pronounced sheeting which is well developed in the quarry of the Presbrey-Leland Company, Incorporated (Fig. 2). This structure consists principally of concentric, steeply dipping joints following

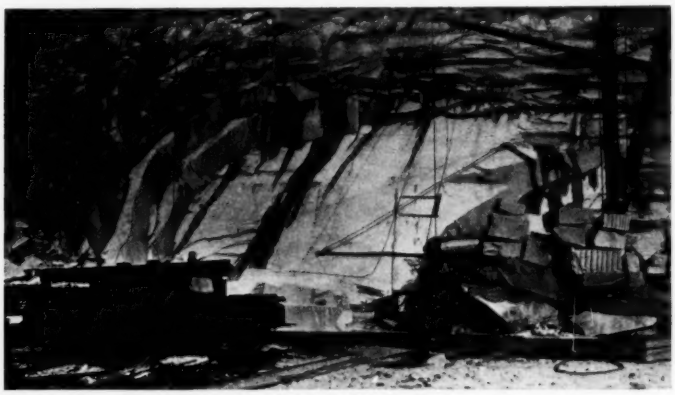

FIG. 2.—The Presbrey-Leland Quarry near West Dummerston, Vermont, showing the well-developed sheeting which decreases with depth. The number of cross joints also decreases with depth, which fact makes it possible to obtain larger blocks.

the outline of the intrusive. In the quarries one finds that the thickness of the sheets or spacing of the joints increases with depth and eventually the jointing will probably disappear. The southern part of the intrusive is quite massive, and jointing in the form of sheeting, or otherwise, is relatively inconspicuous although it is present to a minor degree.

The texture of the rock, examined megascopically, is of relatively uniform medium grain with the exception of a few small areas where it is somewhat finer or coarser. The rock in the southern part of the intrusion is, however, slightly coarser in texture than that in the northern part and is due, no doubt, to its having crystallized at

slightly greater depth, as it is now at a lower elevation. It is also darker in color because of a slightly higher percentage of biotite. In very restricted areas the biotite may become an important constituent.

Pegmatites, aplite dikes, and quartz veins occur very sparingly within the intrusive. They vary in width from less than an inch to about 6 inches. The texture of the pegmatites is medium-grained since the size of the crystals is seldom more than an inch or two.

Collecting the samples.—Representative specimens were collected at many outcrops around the periphery and at as nearly regular intervals along traverses toward the center of the massif as outcrops would permit. Only the freshest material obtainable was sectioned. The positions from which these were collected, together with the mineral percentages, were plotted on a map in order to detect any zoning of mineral species around the contacts. The plot showed a high degree of uniformity throughout the entire area except for a slightly higher percentage of biotite in the southern part of the igneous mass.

MINERAL ANALYSIS AND CLASSIFICATION

A Leitz six-spindle integrating stage was used in the quantitative mineral determination of the sections examined, and the results, in so far as the method is concerned, are considered to be accurate[4] to about 1 per cent. The average results of analyses of 82 slides of the intrusive, 10 slides of sills and dikes, and 1 slide of the aplite together with an analysis reported by Maynard[5] are given in Table 1.

From the average percentage mineral composition of the intrusive mass and of the sills it is clear that these differ very little from one another mineralogically. However, the sills are slightly richer in quartz and epidote and contain less biotite and magnetite than the main mass. The increased amount of silica in the small aplite dikes is undoubtedly due to the process of fractionation which removed the plagioclase feldspar, the predominant mineral to crystallize early, and thereby produced a residual solution enriched in quartz.

[4] E. S. Larsen and F. S. Miller, "The Rosiwal Method and the Modal Determination of Rocks," *Amer. Mineral.*, Vol. XX, No. 4 (1935), pp. 266–73.

[5] Maynard, *op. cit.*, p. 147.

Since the stock, sills, and dikes do not differ greatly in composition, they all fall into the same classification, and according to that of Johannsen the rock is Family 127 or a leucogranodiorite.

Dale[6] and Maynard[7] have both made studies of this intrusion and differ chiefly in their classification. Maynard determined the rock to be a leucogranodiorite as the result of a quantitative study of the minerals. Dale classified it as a quartz monzonite. The writer believes that the amount of orthoclase present is less than that determined by either of these investigators and is probably less than 3 per cent, for in none of the slides could more than a few grains be found and in many of the sections orthoclase is absent.

TABLE 1

AVERAGE PERCENTAGE MINERAL COMPOSITION

Mineral	Main Part of Intrusive—82 Sections	Sills and Dikes in Country Rock—10 Sections	An Aplite within the Massif—One Section	Analysis of Maynard
Quartz	29.74	32.03	40.46	23.88
Microcline	12.79	11.59	13.74	12.19
Orthoclase	little	little	little	7.18
Albite-oligoclase	50.80	50.92	41.49	49.98
Biotite	1.90	little	none	4.28
Muscovite	4.60	4.46	4.43	2.29
Magnetite	little	little	none	none
Garnet	little	little	none	none
Epidote	little	little	none	little

PETROGRAPHIC DESCRIPTION OF THE LEUCOGRANODIORITE

In thin section the leucogranodiorite shows a hypidiomorphic granular texture, the grain size averaging up to about 0.6 mm. in the southern part of the area while in the northern part it is somewhat finer grained.

The minerals are quite fresh with the exception of the oligoclase in which the central part is more or less altered and thus provides an excellent criterion for distinguishing untwinned plagioclase from the potash feldspars which are free from zoned alteration. Pressure effects are clearly shown by the pronounced wavy extinction of quartz

[6] *Loc. cit.* [7] *Op. cit.*, p. 146.

and to a lesser degree of the feldspars and the micas. In the sills, and to some extent in the dikes, the minerals are thoroughly granulated, and in the former the fragments are often drawn out like flow lines (Fig. 3).

A few small aplitic areas or zones occur in the main intrusion. The texture of this phase of the leucogranodiorite consists of a fine and uniform grained aggregate of quartz, microcline, oligoclase, and

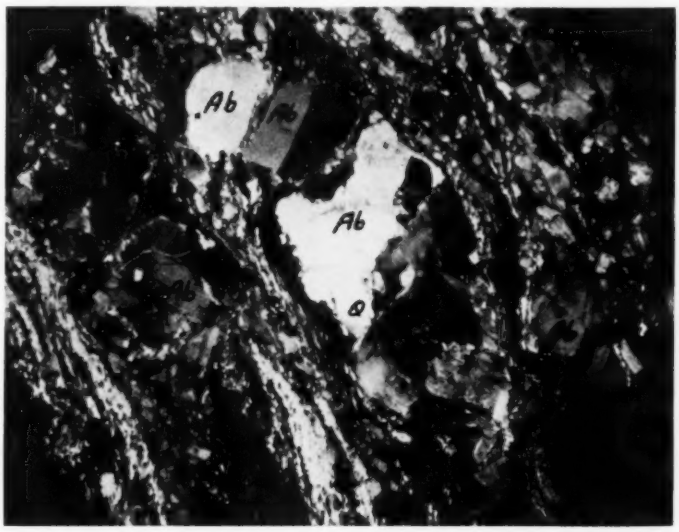

FIG. 3.—Photomicrograph of material from a sill showing the extreme phase of crushing. ×10.5.

muscovite. The aplite escaped the disturbances which produced such pronounced cataclastic effects in the sills and to a lesser degree in the dikes surrounding the main igneous mass. It is probable that there was some movement which caused the crushing within the sills before the magma was completely crystallized

Quartz.—Quartz is always rather clear and colorless, occurring as anhedral individual grains and also interstitially between oligoclase and microcline. In many cases it has granulated boundaries and sometimes occurs as crushed streaks within larger quartz grains; and

also as rounded or curved elongated inclusions in oligoclase as vermicular intergrowths. Wavy extinction, inclusions of sericite, needle-like apatite crystals, and sometimes minute unidentified hairlike crystals occur in the quartz. Fluid and dustlike inclusions are present in zigzag bands across some grains.

Albite-oligoclase.—The albite-oligoclase in some cases is subhedral, colorless, twinned on albite, pericline, and carlsbad laws, and often

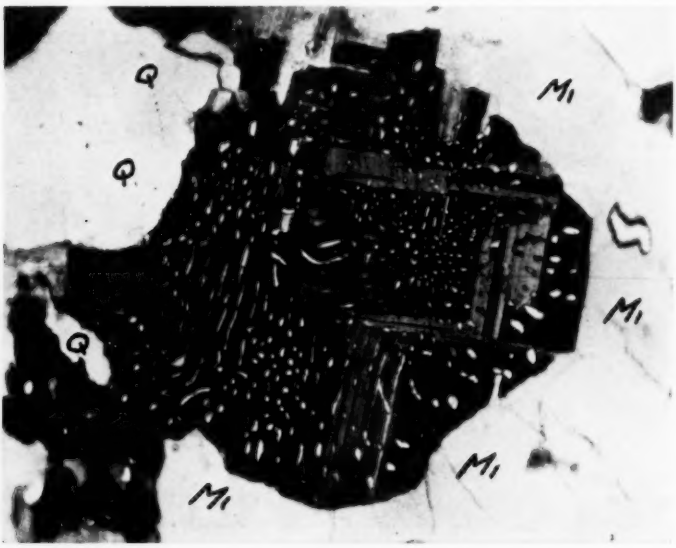

FIG. 4.—Myrmekite grain having both albite and pericline twinning, which is not common for this intergrowth is usually without twinning. The grain is nearly surrounded by extremely faintly and locally twinned microcline. ×66.6.

contains inclusions of muscovite and sericite with sometimes chlorite, magnetite, and epidote. Practically all of the grains of this mineral show either alteration or composition zoning. Myrmekite an intergrowth of quartz and plagioclase feldspar like that described by Sederholm[8] as occurring in the granitic rocks in Finland, is rather common and sometimes exceptionally well developed (Fig. 4). Frequently the plagioclase is untwinned, in which event its identi-

[8] J. J. Sederholm, "On Synantetic Minerals and Related Phenomena," *Bull. de la Comm. Geol. 48* (Finland, 1916).

fication could readily be made by checking its index against the Canada balsam or by means of the presence of zoning and alterations of the central portion of the grains which differ from, or are absent in, the potash feldspars. A few grains of albite-oligoclase show a zoned extinction parallel to the outline of the grain, the interior having the greater extinction angle. This indicates an original core probably much richer in the anorthite molecule. Since anorthite-

FIG. 5.—Photomicrograph showing localization of well-developed twinning in microcline around quartz grains and along the contacts. These grains were no doubt originally orthoclase. A decided wavy extinction is also present. ×27.5.

rich plagioclase alters much more readily than albite, the greater degree of alteration of the center of the grains is thus readily explained and helps materially to confirm the identity of the untwinned feldspar as plagioclase.

Orthoclase-microcline.—Orthoclase, the high temperature form of the potash feldspar, is rather uncommon in the leucogranodiorite. Many sections of the rock contain grains in which microcline twinning is partially developed (Fig. 5). Whenever this twinning is present, even in small areas, the grain is considered to be the low

temperature form. The partially twinned grains probably represent original orthoclase which became sufficiently unstable during the end stages of crystallization to revert to microcline.

The potash feldspars contain fewer inclusions than the albite-oligoclase and when present are more uniformly distributed. They are also considerably fresher and lack zoning, but they show pronounced wavy extinction. These two groups of feldspars were readily separated on the integrating stage by the differences in their indices of refraction and by the characteristic arrangement of inclusions in each.

Biotite.—Biotite is anhedral, occurs as laths, elongated grains, or shreds in cross section, and is often closely associated with muscovite. Alteration, shredding, strain shadows, and shearing are often present. Pleochroic haloes around zircon inclusions are present in some of the grains.

Muscovite.—Muscovite occurs in the same relationships as does biotite, often intimately associated with it and sometimes in parallel growth.

Accessory and secondary minerals.—Magnetite is scattered through the rock as minute particles, seldom in crystal form. Hematite occurs sparingly as dustlike grains or sometimes as larger, irregularly shaped masses; it is red-brown in color and occurs in oligoclase, quartz, and around the edges of garnets. Apatite as minute needles or in hexagonal outlines is always associated with quartz and feldspar. Zircon occurs entirely within the biotite. Sericite is present as an alteration product in the central part of the albite-oligoclase crystals. Epidote occurs as aggregates associated with sericite in the plagioclase and is probably a secondary mineral derived from the feldspar.

PETROGRAPHIC DESCRIPTION OF SOME OF THE SCHIST STRATA

The schist strata are too variable to be described in detail. The sections examined show characteristic parallel orientation of the constituent minerals which vary considerably in amount in the different specimens. In some cases quartz predominates, in others the percentage of calcite is equal to or greater than that of the quartz. The percentages of biotite, zoisite, and muscovite also vary

greatly in different specimens, and in some instances the amphiboles, chiefly actinolite, predominate. The mineral analyses of four of these specimens are given in Table 2.

TABLE 2

PERCENTAGE MINERAL COMPOSITION OF SOME SCHIST SECTIONS

MINERAL	SLIDE NUMBERS			
	5	6	28	29
Quartz	68.92	69.52	29.30	42.10
Biotite	14.82	13.14	none	none
Zoisite	11.69	14.54	none	none
Oligoclase	4.69	none	none	none
Magnetite	little	little	3.82	2.69
Pyrite	little	little	little	none
Calcite	little	1.64	62.60	47.80
Muscovite	little	little	3.79	7.42

ORIGIN OF THE LEUCOGRANODIORITE AND APOPHYSES

The sediments surrounding the intrusive show only a minor amount of disturbance; at the north and south, as well as along the sides, apophyses penetrate the country rock as sills, dikes, and lit-par-lit injections. Within the crescent a large xenolith occurs which retains its original north and south direction but is injected with many small dikes and sills. These features suggest that the magma entered the sediments in a somewhat orderly manner, gradually assimilating or stoping its way into the overlying country rock chiefly by penetrating the sediments along their bedding planes. The character of the adjacent sediments was not noticeably altered by the incoming magma because these were evidently metamorphosed prior to the injection and hence were quite stable under the physico-chemical conditions existing along the contacts at the time the magma was injected. The extent of solution of the inclosing sediments depended entirely upon the degree of superheat of the magma.

CONCLUSIONS

The average percentage mineral composition of 82 slides from the Black Mountain leucogranodiorite is: quartz, 29.74; microcline, 12.79; albite-oligoclase, 50.80; biotite, 1.90; muscovite, 4.60. The

form of the intrusive is that of a stock having a complement of sills and dikes penetrating the country rock.

Orthoclase occurs at present as a minor accessory mineral. Earlier, no doubt, the percentage was somewhat higher, but most of it has apparently reverted to the lower temperature form, microcline, as the result of rather high differential pressure in an environment where orthoclase is less stable than its paramorph, microcline. The magma became more sodic as crystallization progressed, as is shown by the more albitic borders of the plagioclase feldspars.

ACKNOWLEDGMENTS.—The author gratefully acknowledges indebtedness to the members of the geology department of Smith College—especially to Professor B. M. Shaub—for numerous helpful suggestions during the progress of the field and laboratory work and the preparation of the manuscript.

THE RECENT ADVANCE OF BLACK RAPIDS GLACIER, ALASKA

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ABSTRACT

Excess precipitation averaging perhaps 17 per cent from 1928 through 1931, over a large collecting ground with a narrow discharge lane, may account for the remarkable average advance of about 115 feet per day which the front of the Black Rapids glacier appears to have maintained between September, 1936, and February, 1937, resulting in an increase of about 5 per cent of ice-covered area.

Black Rapids glacier is a large ice sheet, perhaps 25 miles long, the result of the merging of eleven or more smaller valley glaciers which originate in the Alaskan Range on the southeast slope of Mount Hayes and on the north flanks of parallel spurs to the south. It descends eastward to the valley of the Delta River at Rapids Roadhouse on the Richardson Highway, 233 miles by road north of Valdez and 138 miles by road south of Fairbanks. A topographic map of this area is included in *Bulletin 498* and *Bulletin 501* of the United States Geological Survey, published in 1912.[1]

According to the maps in these bulletins, the lower limit of the glacier in 1912 was the west edge of the Delta River Valley, with a terminal moraine extending nearly a mile east of the ice margin out over the valley floor (Fig. 1). During the summers of 1935 and 1936 the lower margin of the glacier occupied a position approximately 3 miles west of the Delta Valley proper, indicating a recession of about that distance since 1912. In fact, from 1935 to 1936 the lower margin retreated westward, perhaps ½ mile according to several observations by different people, including myself, who had noted its position from the Richardson Highway, some distance to the east. Several hunters and prospectors reported that during the summer of 1936 its lower edge could be crossed with ease. Late in August, 1936, the

[1] Fred H. Moffit, "Headwater Regions of Gulkana and Susitna Rivers, Alaska," *U.S. Geol. Surv. Bull. 498* (1912); Stephen R. Capps, "The Bonnifield Regions, Alaska," *U.S. Geol. Surv. Bull. 501* (1912).

ice front visible from the Richardson Highway was low and about 3 miles distant from the Delta River Valley.

About the middle of December, 1936, the Fairbanks Weather Bureau informed me that the people living at Rapids Roadhouse were alarmed at the sudden advance of Black Rapids glacier, and asked if it were probable or possible that the ice would advance across the Delta River. I called their attention to the recent morainal material in the Delta Valley, showing that in the very recent

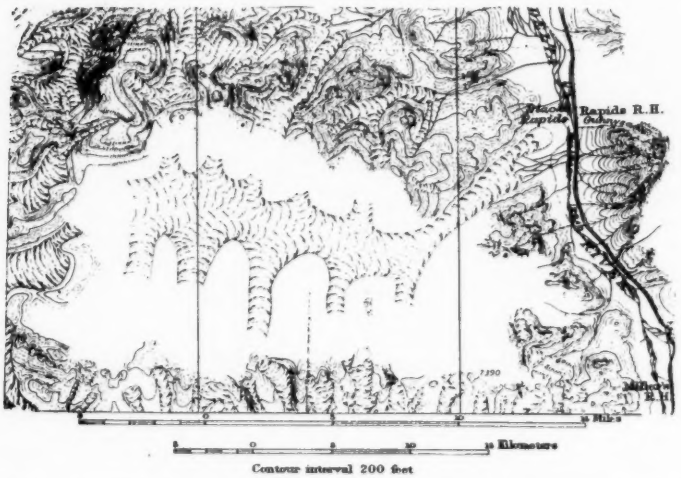

FIG. 1.—Map of Black Rapids glacier and the large catchment area which feeds it. From *U.S. Geol. Surv. Bull. 498.*

past the ice had been out beyond the mouth of the tributary valley, but that I had not seen any evidence of this glacier having crossed the Delta River Valley. To me it appeared possible but not probable. Again in January, 1937, I received a similar message stating that the ice wall was still advancing and was nearly out of the tributary valley. Late in February, 1937, I made an airplane trip to the locality and spent some time in front of the glacier. Because of turbulent air conditions over the glacier, and because no landing facilities were available there at that time, the examination was brief and marginal. The accompanying illustrations were taken from an

elevation of about 1,000–1,500 feet above the ground, and about ½ mile from the ice front. The steep lateral walls, the abrupt front, the

FIG. 2.—View westward up Black Rapids glacier showing rough surface, abrupt front wall, and older moraine still uncovered. Mountain in background about 10 miles distant. February, 1937.

FIG. 3.—North side of glacier near front. Right side of picture shows moraine of a former ice advance. February, 1937.

extremely rough surface of the ice, and the still uncovered moraine of an earlier ice advance are well shown (Figs. 2, 3, and 4).

During the winter of 1936–37 the margin has thus readvanced about 3 miles eastward down the tributary valley, and again is at the edge of the Delta River Valley. Unfortunately the exact time when this remarkable advance began is not known. Early winter snow closes part of the Richardson Highway about September 15 and thus initiates a cooler period of slower ice wastage and of greater ice augmentation. Hence, if we consider that this ice front began its unusually rapid advance late in September or early in October, and that the advance amounted to 3 miles (an estimate, but apparently

FIG. 4.—Fractured surface of Black Rapids glacier. The abrupt front wall is about 400 feet high.

a reasonable one), the average forward movement of the front per day would approximate 115 feet—a figure somewhat in excess of one other reported instance[2] of 100 feet a day. Between February and June of 1937 local observers reported advances as great as 250 feet per day for the Black Rapids glacier, but crumbling and flattening of the front slopes may account for some of these phenomenal figures. At any rate there has been little if any appreciable advance of the ice front as a whole between late February and June.

Relatively stagnant ice, covered with morainal debris, extended for an unknown distance forward from the visible ice front prior to

[2] R. T. Chamberlin, personal communication regarding a Greenland glacier.

this recent advance. The fresher or newer ice from above has plowed forward through this stagnant material and shows a wedge of light-colored ice apexing near the front margin, and is bordered on both sides by ice which is much darker in color due to the rock waste which had been accumulating as a recessional moraine. The lower margin of this glacier consists of ice conspicuously fractured and crushed, apparently the result of higher ice plowing down through stagnant ice rather than riding over a smooth floor, and so broken only by crevasses (Figs. 5 and 6). Ever since January of this year the

FIG. 5.—Left ice foreground shows old ice and recessional moraine plowed forward in crushed ice. February, 1937.

entire surface of the frontal portion has presented an extremely rough and pinnacled appearance, indicating unusual crushing and fragmentation. Airplane photographs show that this condition extends at least 5 or 6 miles back from the lower margin. The pictures taken in February show the ice with a snow covering, but when seen later in June the lower margin showed much rock debris mixed with the ice and covering the glacier front. Estimates of the height of the ice front vary from 200 feet to 500 feet, and its width is that of the tributary valley floor, about 2 miles. In February the side margins were nearly vertical walls, and the front was steep, about 50° from the horizontal. At that time the ice front was within $\frac{1}{2}$–$\frac{3}{4}$ of a

mile of the extreme edge of a very recent terminal moraine showing typical kame and kettle topography.

The topographic map does not show in detail the character of the collecting area for this glacier, but indicates that the area of snowfall tributary to it must be in the neighborhood of 100–150 square miles, perhaps more. It is therefore very large whereas the lane of discharge is relatively very narrow.

FIG. 6.—Shows the complete filling of the valley floor, the rough surface of the ice, and the rock and mud covering near the side and end margin. Central portion appears lighter in color. June, 1937.

Obviously the extent of any snow field and the glacier or glaciers arising from it will depend upon such factors as amount and rate of snowfall, melting and evaporation, topographic details which determine the area of the catchment basin and affect the depth of accumulating ice, topographic slope, slope of the upper surface of the ice, and the character of the discharge lanes.

There are no records of precipitation in this basin, but a fairly complete record of snow and rainfall since 1904 is available for Fairbanks, about 100 miles to the north. Although correlation with Fairbanks might be considered as of long range, a study of that record seems to give a reason for the advance of the glacier. Snowfall on the south slope of the Alaskan Range should greatly exceed

that in the Tanana Valley to the north, but some parallelism between the two is likely.

The average annual precipitation for Fairbanks between January 1, 1905, and December 31, 1936, was 11.805 inches. Replotting this by year periods beginning August 1, and closing the year on July 31, the yearly average is 11.946 inches. This latter year period fits best with the available observations on ice advance and recession.[3]

The three-year periods terminating July 31, 1908, averaged 14.43 inches, or an excess of about 2.5 inches year per. Then followed five years of less than average precipitation, 9.48 inches, and yet the heavier snowfall from 1906 to 1908 managed to bring the ice margin to the edge of the Delta River Valley by 1912, and about where it now is, in spite of the lesser precipitation from 1908 through 1913. From the precipitation record of the next nine years, 1914–22, the annual fall averaged 12.39 inches, which period was followed by six low years, 1923–28, averaging 9.63 inches. According to this, some retreat of the margin after 1912 might have been followed by a readvance about 1926, but of this I have no authentic information.

Following the deficiency period terminating in 1928 came eight years (up to August 1, 1936) with an average annual fall of 12.85 inches, or an excess of 0.9 of an inch per year. The first four years of this period averaged 13.96 inches, an excess of 2 inches per year. The last four years averaged 11.75 inches per year, a deficiency of 0.2 of an inch per year. Perhaps now we are just noting the results of the large excess of snowfall from 1929 to 1932, after a four-year interval.

Various comments, most of them of little value, have been made upon the oscillation of this particular ice front. In view of the large collecting field and the precipitous valley walls which protect the snow from very effective solar evaporation or melting, I am convinced that the striking oscillations of the lower margin of this glacier are related to differences in snowfall, and that the excesses or

[3] A complete record by months of the Fairbanks area from October, 1904, through January, 1937, was furnished me by Mr. Frost, assistant meteorologist of the Weather Bureau, stationed at Fairbanks.

deficiencies may require about four years to be felt at the lower ice front.

The following facts appear pertinent and adequate to account for the recent abrupt and rapid advance of the Black Rapids glacier: (*a*) the collecting field is large, about 125 square miles; (*b*) this east-west tributary valley which functions as a discharge lane is narrow at the lower end, and is bordered by a high mountain wall on the south which effectively protects the ice from the sun which is low in the sky in this latitude; (*c*) frozen ground below the ice sheet prevents much loss of water downward into the rock floor. Hence the disposal of snow and rain which fall within this catchment basin is almost entirely by evaporation and runoff, largely the latter. When evaporation lags, an increase in runoff may be expected. Such a readjustment during the winter season would normally produce an advance of the ice at the lower margin, which in this case is through a narrow outlet of less than 2 miles in width. Crushing and grinding of the ice in such restricted areas, and where it is forced through stagnant ice at the foot, would be expected, and suggests movement of a solid rather than of a plastic mass. Although we have had numerous earthquakes reported in the Tanana Valley during 1936–37, and the past winter has seen excessive snowfall in Fairbanks, neither of these appears to have been a factor of importance in producing the present ice extension. Accumulation of excess snowfall over several years, with a gradual increase of the ice below, is entirely adequate to explain this recent advance which appears to have been due to the greater snowfall indicated since 1928, which lasted for at least four years, and was then followed by four more years of nearly normal precipitation before the last season.

In a study of this character it may be of interest to attempt an estimate of the increase in snowfall necessary for such an ice extension. Because there is not yet available any record of precipitation for this particular area, such an estimate may be suggestive only. The ice extension appears to cover an area approximately 3 by $1\frac{3}{4}$ miles, and to an average depth of about 300 feet. This is equivalent to 0.298 cubic miles. If such a volume of glacial ice were spread over the entire ice-covered area of about 125 square miles, it would make a layer about 12.58 feet thick. If this amount were due to excess ac-

cumulation over an eight-year period, the annual increment would average 1.57 feet. How much glacial ice may result from one foot of snow is uncertain. Perhaps 3 inches is a fair estimate. If so, the average snow excess per year would be 1.57 feet $\div \frac{3}{12}$, or 6.29 feet for the last eight years.

Increased snowfall over the catchment area, especially in the higher portions, means a greater thickness of the ice field with a correspondingly steeper slope of the ice surface. Such steepening of the ice slope would probably reach a maximum at the higher altitudes where the rock slopes are also steepest. The thickened ice sheet would offer little or no additional surface for wastage by evaporation or melting except in the marginal extension at the lower end. In the case of Black Rapids glacier, the abrupt advance of the lower margin with little or no apparent warning symptoms is noteworthy. When the ice margin reaches the Delta Valley, wastage by evaporation and melting is greatly accelerated because of prevailing air currents and solar exposure, and hence the western edge of the Delta Valley would appear to be a zone beyond which such ice extension would rarely if ever advance without striking meteorological variations.

THE RATE OF ADJUSTMENT OF CALCIUM BICARBONATE SOLUTIONS

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ABSTRACT

A theoretical and experimental study of the rate of adjustment of calcium bicarbonate solutions indicates that evaporation is not an important rate-determining factor in the process. The observed catalytic effect of calcite is quite general for any added surface.

E. Steidtmann,[1] has raised the question: "Does evaporation increase the rate of adjustment of supersaturated $CaCO_3$ solutions?" The following considerations and observations are offered in an endeavor to present more convincing evidence upon this important geological question.

The rate of adjustment of exposed calcium bicarbonate solutions will depend upon the following primary factors: (1) rate of loss of CO_2; (2) rate of loss of water; (3) rate of crystallization of $CaCO_3$ formed as a consequence of 1 and 2.

Considering the third factor first, we should expect it to be rapid in comparison with processes 1 and 2. It is possible that certain inhibiting mechanisms could operate in such a manner as to be interpreted as the operation of factor 3. The formation of supersaturated $CaCO_3$ solutions as such (as distinct from the formation of $Ca(HCO_3)_2$), supersaturation of CO_2, or the formation of colloidal $CaCO_3$ are such mechanisms. Any simultaneous supersaturation of the two products of decomposition is irrelevant because such a case would be equivalent to the original $Ca(HCO_3)_2$ solution. Any individual supersaturation is unlikely, in nature or under the conditions of the experiments to be described, because of the agitation and ever present nuclei.

The possibility of colloid formation is worthy of detailed consideration. The slow rate of formation of $CaCO_3$ from $Ca(HCO_3)_2$ solution will discourage the formation of the colloidal state.[2] This

[1] "Travertine-depositing Waters near Lexington, Virginia," *Jour. Geol.*, Vol. XLIV (1936), p. 193.

[2] P. von Weimarn, "Grundzüge der Dispersoidchemie," Dresden: Theodor Steinkopff (1911).

was confirmed, as suggested by Steidtmann, by determining the total CO_2 of partially decomposed solutions. In all cases the results agreed splendidly with the titration values. More convincing than this and other evidence is the fact that upon examination of partially decomposed solutions under dark-ground illumination from a Bausch and Lomb paraboloid condenser the solutions were all optically void.

The rate of adjustment of a $Ca(HCO_3)_2$ solution is therefore determined by the rate of loss of CO_2 and of water. The controlling factors in the loss of these two materials are their diffusion rates, which are best set up in the form of the Whitman two-film theory.[3] The results are:

$$\frac{dW_{CO_2}}{d\theta} = k_{CO_2}(C_1 - C_g) = k_{CO_2} \cdot S$$

$$\frac{dW_{H_2O}}{d\theta} = k_{H_2O}(H_s - H_g)$$

where $dW/d\theta$ is the rate of loss of the specified material; the k's are overall coefficients, determined by the diffusion constants and the particular conditions; C_1 and C_g are the concentrations of CO_2 in the liquid and gas phases, respectively; and $C_1 - C_g$ is the supersaturation of the solution. H_s and H_g are the humidities of saturated and actual air.

The loss in total dissolved $CaCO_3$ [$CaCO_3 + Ca(HCO_3)_2 + CO_2$ (potential $CaCO_3$)] from a solution as it loses these two determining constituents will be

$$\frac{dW}{d\theta} = K'_{CO_2} \cdot S + K'_{H_2O}(H_s - H_g) ,$$

where the change from k to K' takes care of the effect of each of these constituents on the equilibrium: $Ca(HCO_3)_2 \rightleftarrows CaCO_3 + CO_2 + H_2O$. K'_{H_2O} is composed of the product of two factors; the diffusion constant k_{H_2O}, and a factor determined by the activity of

[3] W. G. Whitman, "The Two-Film Theory of Gas Absorption," *Chem. & Met. Eng. 29* (1923), p. 147. W. K. Lewis and W. G. Whitman, "Principles of Gas Absorption," *Jour. Ind. & Eng. Chem. 16* (1924), p. 1215.

the water in the progressively decomposing solution. For dilute solutions the latter remains practically constant.

An attempt to estimate the ratio of the two K's from the complete form of the foregoing equation, and the scant data available, gave a value of 85; which, for practical purposes, would justify ignoring K'_{H_2O}.

The individual constants may be evaluated as follows: If a $Ca(HCO_3)_2$ solution is allowed to lose only CO_2, in a water-saturated atmosphere, $H_s - H_g = 0$. Changing over to concentration by means of $dW/V = -dC$, integrating, and evaluating the constant by $C = C_0$ when $\theta = 0$,

$$K_{CO_2} \cdot \theta = \log \frac{C_0 - C_s}{C - C_s}.$$

K_{H_2O} may be evaluated by losing H_2O and no CO_2:

$$K_{H_2O} \cdot \theta = \frac{C_0 - C}{H_s - H_g},$$

from whence it is obvious that $K_{H_2O} = 0$, since $C_0 - C$ sensibly equals 0.

EXPERIMENTS AND RESULTS

Calcium bicarbonate solutions were allowed to decompose in a closed system provided with wet- and dry-bulb thermometers, manometer, sampling tube, perforated bubbler, and flowmeter. The bubbler was fixed in position; and in all runs the same amount of solution, with excess $CaCO_3$, was used. The required atmospheres were obtained by previously passing the air through proper liquids. The analytical procedures were the standard water methods of the American Public Health Association. The use of a filter aid and Whatman paper No. 42 eliminated all irregularities due to fines.

Preliminary experiments indicated that even at room temperatures the rate of adjustment of the solutions was inconveniently slow. Increased agitation, although accelerating the change, did not bring it within convenient range; therefore the temperature was raised to $34 \pm \frac{1}{4}°$ C. In all runs the system was brought to a steady

state for at least one-half hour before taking zero time. The concentrations corresponding were usually 200–250 ppm.

Representative results for two sets of comparative runs are given in Figures 1 and 2. The conditions within each set differ only in the extremes of humidity of the bubbled air. The temperature of set Nos. 6–8 was about $\frac{3}{4}°$ lower than the others. The exact parallelism of comparable curves in Figure 1 demonstrates that even the maximum variation in humidity of the adjusting air does not affect the

TABLE 1

Run No.	Temperature in Degrees Centigrade	Entering Air		Air Flow, Cc/Min.	K_{CO_2}
		Per Cent of Humidity	Per Cent of CO_2		
3	20	100	0.04	20	0.018
4	17	100	.04	320	.015
4A	18	100	.00	320	.013
5	34	100	.04	105	.061
8	33.3	100	.04	110	.054
9	34	100	.04	108	.059
10	34	100	.04	110	.058
6	33.3	0	.04	106	.057
7	34	0	.04	108	.058
7A	34	0	.04	110	.058
11	34	100	.00	108	.058
12	34	0	.00	112	.055
13	34	42	0.04	110	0.059

Average K_{CO_2} at 34° and 100 per cent Humidity = 0.058
Average K_{CO_2} at 34° and 0 per cent Humidity = 0.057

rate of decomposition of these solutions. This is shown better in Figure 2, where a straight line would result for No. 6 and No. 7 only if K_{H_2O} was negligible compared to K_{CO_2}. The experimental results are summarized, by means of K_{CO_2}, in Table 1.

The constancy of K indicates that even the maximum rate of loss of water by evaporation has a negligible effect upon the rate of adjustment of the solutions. As equilibrium is approached this factor will begin to exert an effect upon the amount of $CaCO_3$ deposited from a given solution; and it can be calculated from the observation that for the particular conditions employed, 8 g. of water evaporated per hour. The evaporation estimated from empirical equations is

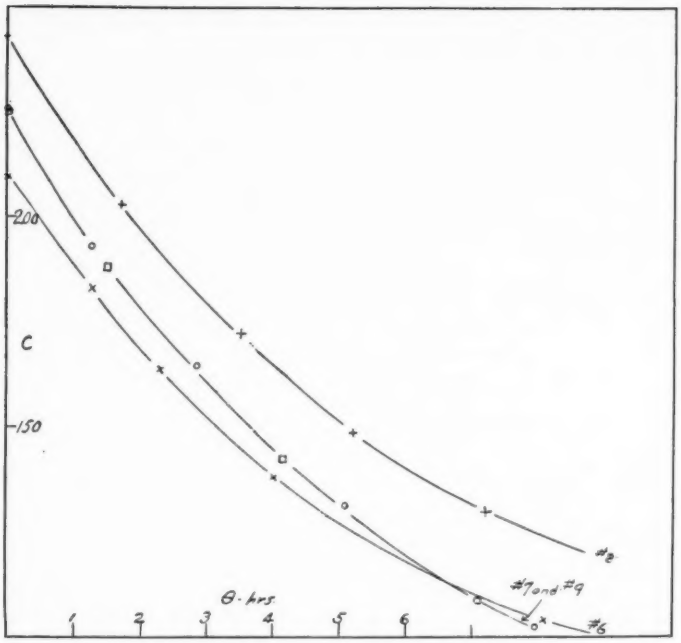

FIG. 1.—Rate of decomposition of $Ca(HCO_3)_2$

× No. 6—air— o per cent humidity
\+ 8— —100
o 7— — o
□ 9— —100

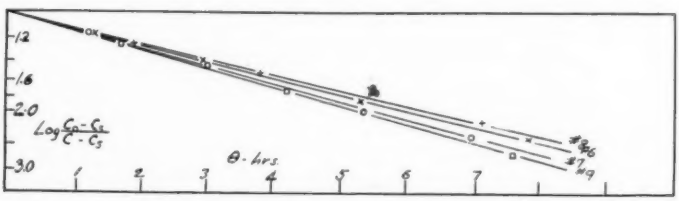

FIG. 2.—Rate of decomposition of $Ca(HCO_3)_2$

× No. 6—air— o per cent humidity
\+ 8— —100
o 7— — o
□ 9— —100

17 g.;[4] 13.0 g.[5] The value of K_{CO_2} remains constant down to about $S = 20$ ppm. Below this value there is an irregular variation, most likely due to analytical uncertainty. The observed end-points agreed with those calculated from the equilibrium constant for this system.

The form of the final rate-determining expression suggests that the rate of adjustment of similar bicarbonate solutions should be independent of the cation. To test this suggestion similar solutions of calcium, magnesium, strontium, manganese, and zinc bicarbonates were allowed to decompose under identical conditions. All decomposed in a first-order manner with the following constants: 0.052, 0.054, 0.037, 0.054, and 0.047, respectively. This same behavior has been observed for the absorption of CO_2 by alkaline solutions.[6] Supersaturated aqueous solutions of CO_2 did not decompose at all in the same manner.[7]

A constant excess of $CaCO_3$ was present in all experiments, in order to fix the catalytic effect of this material. A few decompositions in open beakers, with agitation, and in the presence of fine, insoluble materials (such as marble, quartz, sulphur, manganese dioxide, barite, clay, etc.) indicate that all accelerate the decomposition relative to the case of no initial excess $CaCO_3$. To a certain extent the catalytic effect appears to be a function only of the amount of added surface. It must be regarded, therefore, as the physical effect of added surface in assisting the formation of crystalline $CaCO_3$ and the escape of CO_2.

CONCLUSIONS

Loss of water by evaporation has no significant effect upon the rate of adjustment of calcium bicarbonate solutions.

The catalytic effect of calcite is not specific, but is merely the effect due to enhanced surface in assisting adjustment.

[4] J. W. Hinckley, "The General Problem of Evaporation," *Jour. Soc. Chem. Ind.-London 41* (1922), p. 242 Transactions.

[5] M. Lurie and N. Michailoff, "Evaporation from Free Water Surface," *Jour. Ind & Eng. Chem. 28* (1936), p. 345.

[6] L. B. Hitchcock, "Mechanism of Gas-Liquid Reaction," *Jour. Ind. & Eng. Chem. 29* (1937), p. 302.

[7] See also, "T. K. Sherwood, F. C. Oraemel, and N. E. Ruckman, *ibid.* (1937), p. 282.

GLACIAL LAKES PONASK AND SACHIGO DISTRICT OF KENORA (PATRICIA PORTION), ONTARIO[1]

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ABSTRACT

Interpretation of glacial field data in the vicinity of the Manitoba-Ontario boundary, northwest Patricia, Ontario, leads the writer to postulate the existence of an interlobate moraine and the former existence of two glacial lakes to be known as Ponask and Sachigo.

In the course of a geological reconnaissance of the Stull Lake–Echoing River–Sachigo Lake area for the Ontario Department of Mines during the 1936 field season the writer observed an interesting association of varved clays, glacial striae, and morainic deposits which throw additional light on the retreat of the continental glaciers in the neighborhood of the Ontario-Manitoba boundary between latitudes 53°30′ N. and 55° N.

Varved clays have a widespread distribution in a region including Little Sachigo, Sachigo, and Ponask lakes. In this region the clays are essentially flat-lying and extend for an east-west distance of 20 miles and a known north-south distance of 30 miles. Bisecting this belt of clays is a moraine ridge which rises 200 feet above the level of the clay plain. This ridge trends approximately north and south and is a marked feature of the landscape of the region. A ridge of related origin is also to be seen south of Echoing Lake.

Glacial striae were found in a number of localities. On Stull Lake the movement was from S. 18° E. to S. 28° E. based on twelve readings. Only two marked exceptions occurred outside the above range, S. 3° E. and S. 43° E. On Ellard Lake the direction of ice movement was S. 50°–58° W. One reading from the south part of Pierce Lake was S. 20° E. On Ponask Lake due south of Stull Lake readings varied from S. 22° E. to S. 25° E. but toward its eastern end the

[1] Published with the permission of the provincial geologist of the Ontario Department of Mines.

direction changes through S. 35° E. at 7 miles to S. 45° E. at 4 miles northwest of the 200-foot high moraine ridge on the Sachigo portage. On Sachigo Lake, east of the moraine ridge, glacial striae indicate a southwest movement, the following directions being found: S. 67° W., S. 64° W., and S. 44° W. West of Ponask Lake directions of S. 2° E.–S. 5° E. were found on Seeber Lake, and S. 0° E. on Sandymac Lake. These readings, combined with those obtained on

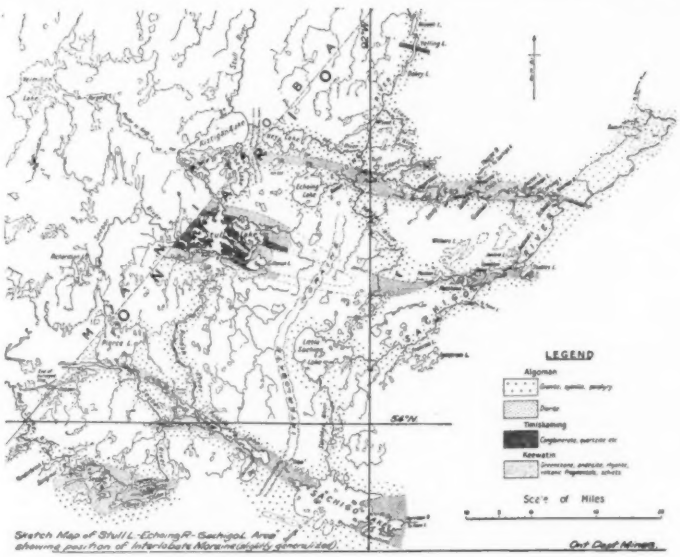

FIG. 1.—Sketch map of Stull Lake–Echoing River–Sachigo Lake area, showing position of interlobate moraine (slightly generalized).

Ponask Lake, suggest the former existence of a deploying ice lobe in this region.

It was noted in the foregoing that on the Sachigo Lake side of the moraine the glacial striae indicated an ice movement in a direction S. 65° W., while on the Stull-Ponask lakes side of the ridge the ice movement was S. 25° E. (average). It is therefore thought that this ridge represents an interlobate moraine formed between an ice lobe occupying the Stull-Ponask lakes area, and another ice lobe covering the area between Sachigo and Little Sachigo lakes. It would appear

that, on the retreat of these lobes, their melt-water was dammed between the lobes and the previously mentioned moraine and that the varved clays known there were deposited in the glacial lakes thus impounded.

The varved clays on Ponask Lake extend at least 8 miles northwest from the interlobate moraine, and 20 miles east from it on Sachigo Lake, although the best exposures lie within the first 12 miles and the clays appear to thicken as the moraine is approached.

TABLE 1

East Side	West Side
Inches	Inches
3.0	...
2.5	...
1.5	1.4
2.0	2.4
1.3	1.6
2.0	2.0
0.5	0.6
0.9	0.8
1.5	1.4
2.4	2.3
2.0	2.0
1.6	1.6
1.7	2.0
1.1	1.3
2.3	2.4
2.3	2.3
1.9	...
1.5	...

On Anchicum Bay, Ponask Lake, varved clays are exposed in a number of cliff sections. Unfortunately, although these cliffs are from 10 to 30 feet in height, the clay has slumped so that in most localities vertical sections can only be seen for a few feet at the base of the cliffs. At localities on the east and west sides of Anchicum Bay, 2 miles apart, the sections given in Table 1 were measured. By correlating the two thinnest varves, the other varves show an excellent correspondence. The summer portion of a varve is gray in color, weathering to buff; the winter portion is a brownish gray, weathering to a deep chocolate-brown color. The upper contact of the winter portion is sharp, but the lower is not so well defined. The thickness

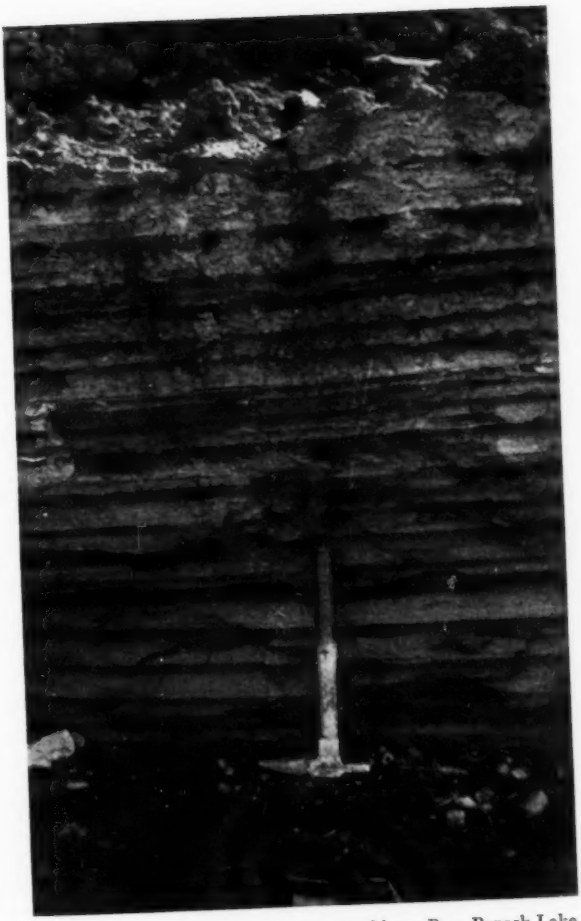

FIG. 2.—Measured section, east side, Anchicum Bay, Ponask Lake

of the winter band varies from 0.2 to 0.7 inches and constitutes approximately one-quarter of the total thickness of a varve.

The foregoing figures indicate that in one case 32 inches of clay were deposited in 18 years, and in the other 24 inches in 14 years. As the total thickness of the clay is believed to be at least 40 feet, the glacial lakes must have existed for 280 years.

Clays showing the same size of varves are well exposed on Sachigo Lake.

On Little Sachigo the varves are much thicker. At a locality along the shore of the southern part of the lake at the base of a slumped 15-foot cliff undisturbed beds show three varves having a thickness of $7\frac{3}{4}$, 6, and 6 inches, respectively. The winter portions of these varves are $1\frac{1}{2}$, $1\frac{1}{4}$, and $1\frac{1}{2}$ inches, respectively.

Bedded clays of possibly a similar type were seen at the east end of Kistigan Lake where they form high banks, and on the east shore of Rorke Lake.

Varved clays similar to the foregoing have been described by Hurst[2] from Angekum, Opasquia, Muskrat Dam, and Sandy lakes, i.e., between latitudes 53°0′ N. and 53°20′ N. or 70 miles southwest of Sachigo Lake. Derry, in his report[3] on the Ontario-Manitoba boundary, states that there is evidence of two advances of the ice sheet in that region, one leaving striae oriented S. 50°–70° W., the other S. 13°–31° W. Both sets were found on Hilton Lake. It seems possible that the first set represents the Sachigo lobe of ice, which has retreated northeastward from the position of Gorman Lake, and the second set the Ponask lobe of ice retreating in a more northerly direction.

Although the area described lies within the district covered at one time by the Patrician glacier, the directions of the glacial striae do not correspond with the movements of that glacier as mapped by Tyrrell.[4] It would appear they are to be correlated with the move-

[2] M. E. Hurst, "Geology of the Area between Favourable Lake and Sandy Lake, District of Kenora (Patricia Portion)," *Ont. Dept. Mines*, Vol. XXXVIII, Part II (1929), p. 68.

[3] D. R. Derry and G. S. MacKenzie, "Geology of the Ontario-Manitoba Boundary" (12th baseline to latitude 54), *ibid.*, Vol. XL, Part II (1931), p. 16.

[4] J. B. Tyrrell, "The Patrician Glacier South of Hudson Bay," *12th Intl. Geol. Cong.*, pp. 523–34.

FIG. 3.—Varved clay, Sachigo Lake

ments of the Keewatin and Labradorean ice sheets or the Labradorean glacier alone.

The work of Upham[5] and Tyrrell[6] has shown that Lake Agassiz was bounded on the north and east by the retreating Keewatin and Labradorean glaciers, respectively; the northeast corner of the lake being at the junction of these two masses. The southern boundary of the Keewatin glacier was approximately at latitude 55° N. and the western boundary of the Labradorean glacier, as followed from south to north, passed from longitude 95° to 96° W. Such being the case it does not seem likely that the interlobate moraine described in the foregoing could have been formed at the boundary line between the Keewatin and Labradorean glaciers as that boundary was to the west and north of the area, unless the Keewatin glacier had a lobe extending a short distance southward along the eastern margin of Lake Agassiz.

As noted previously the directions of glacial striae on the west side of the moraine indicated that the movement of ice was in the direction S. 0° E.–S. 45° E., and on the east side S. 50°–65° W. To account for these converging ice movements the writer suggested in the preceding the existence of two lobes. These lobes, which may be referred to as the Ponask and Sachigo lobes, are believed to be part of the Labradorean glacier. On the retreat of the lobes their meltwater was impounded between them and the interlobate moraine forming, respectively, Ponask and Sachigo glacial lakes. As by this time, with the separation of the Keewatin and the Labradorean glaciers, Lake Agassiz had drained to Hudson Bay, these glacial lakes may be dated as post–Lake Agassiz in age.

ACKNOWLEDGMENT.—The writer wishes to express his thanks to Professor A. P. Coleman, who very kindly read the manuscript and made a number of suggestions.

[5] Warren Upham, "The Glacial Lake Agassiz," *U.S. Geol. Surv. Mono. 25* (1896).

[6] J. B. Tyrrell, "The Glaciation of North Central Canada," *Jour. Geol.*, Vol. VI (1898), p. 158, Pl. VI.

REVIEWS

Structural Evolution of Southern California. By R. D. REED and J. S. HOLLISTER. Tulsa, Okla.: American Association of Petroleum Geologists, 1936. Pp. 157; pl. 8; line drawings 57, tables 6, folded map in colors 1. Cloth bound, $2.00.

The *Structural Evolution of Southern California* is a sequel to the senior author's *Geology of California*, published in 1933. The purpose of the present work is to bring together the published and unpublished facts and theories which have been variously advocated concerning the structural history of Southern California and, in so doing, perhaps to stimulate further thought and theory on the subject.

Southern California is classified into six major tectonic provinces, the geologic history of each being traced by stratigraphic methods. A comparison of these histories reveals that regions of similar basement type have had comparable histories. The tentative hypothesis deduced is given in the authors' words: "Thus the information available at present enables us to state a local orogenic time-law of our own. So far as the post-Franciscan history of Southern California is concerned, tectonically comparable areas seem to have been similarly affected at all times by orogenic and epeirogenic forces."

This correlation of earth movements includes such a very small area that, in itself, it has little bearing on the question of the world-wide contemporaneity of periodic earth movements. However, it suggests a refinement of the present comparisons of histories of larger and more widely separated regions whereby areas chosen for similar tectonic stability, rather than areas of random selection, would be compared. The authors note that any conclusive correlation proving or disproving the hypothesis of diastrophic contemporaneity may be precluded by the factor of the very short time, geologically, required for orogenies. When it is considered that several separate diastrophic episodes have been found to have taken place in California in the time needed for one faunal change, the fear arises that stratigraphic correlations never can be exact enough for diastrophic and volcanic dating.

In their summary, illustrating the observed similarities and dissimilarities of regional histories, the authors have introduced a novel and useful type of chart-diagram.

This book makes no attempt to explain or describe, in terms of their mechanics, the diastrophic movements of Southern California known chiefly by their influence on sedimentation. It is purely a history.

JEAN GRACE

"The Eocene Sediments of Mississippi," by RALPH EARLY GRIM, *Mississippi Geological Survey Bulletin 30*. University, Mississippi, 1936. Pp. 240, figs. 25.

Ralph Early Grim has reconstructed the paleography and sedimentologic history of Mississippi and the southern Appalachian and piedmont belts during the Eocene by petrographic analysis of the sediments deposited in Mississippi at that time. Not only the agents and conditions of deposition but the location and relief of the source areas, the general climatic and botanic environmental condition of weathering, the distance of travel of the sediments, and the stage in their life-cycles of the streams which transported them, are revealed by his graphs and heavy-mineral comparisons. But had the histograms of the Mississippi sediments been on the same scale as the models with which they were compared, a very much greater conviction as to the accuracy of these determinations would have been given the report. Although this is a fundamental theoretical fault, the actual error induced does not, in most cases, seem so great as might be expected.

The southern Appalachian region, with the adjacent piedmont belt, seems to be conclusively indicated as the source of the sediments of the Midway, Wilcox, Claiborne, and Jackson formations, the two areas contributing sediment unequally and in different proportions at different times. The much-debated dating of the southern Appalachian peneplains is brought closer to settlement by this report. The source area of the Midway sediments was in old age, but, rising to a youthful condition by Wilcox time, it was in maturity during the Claiborne deposition and was again in old age in the Jackson. The Midway and Wilcox deposits were part of the huge delta of a great river which must have been the Appalachian River, whose presence, course, and capture has been previously determined by physiographers. The dating of the capture of this river was made possible by Grim's discovery that the Claiborne and Jackson formations, deposited in a littoral and neritic environment, indicate that there was, in central Mississippi, no longer a single great river as a carrier of sediment but rather many small streams.

JEAN GRACE

An Introduction to Historical Geology. By WILLIAM J. MILLER. 4th ed. New York: D. Van Nostrand Co., 1937. Pp. x+475; figs. 375.

The author has included much new material and many new illustrations in this revision of his introductory text. The size and style of the book have been changed to match the companion book, *Physical Geology.*

The introductory chapters dealing with the significance of historical geology, fossils, rock formations, relations of continents and ocean basins, and geologic time have been rather greatly condensed. Many of the illustrations of the new edition that are taken from outside sources are not so clearly reproduced nor so well chosen as might be desired, although the illustrations of fossil forms are very good.

The paleogeography maps are well prepared and agree in general with those of other standard texts. The names of the divisions of geological time are in accord also with standard texts with the exceptions of Taconian (Georgian) for Waucobian and Medinan for Alexandrian. This minor difference in names is not a serious objection. However, the reviewer feels it might have been advisable at least to name the European equivalents of the epochs of the Mesozoic.

The reviewer is very favorably impressed with this introductory text and believes it to be an excellent book to use in colleges where historical geology must be taught in one semester and where the subject must include structural, stratigraphic, and paleontologic geology.

The clear and concise style of the author, the omission of the superfluous names so current in many new texts, the excellent summaries at the close of the various eras, and the general discussion of life of the eras in separate chapters make this a very serviceable and readable textbook for beginning students.

LLOYD W. FISHER

"Stratigraphic Classification of the Pennsylvanian Rocks of Kansas," by RAYMOND C. MOORE, *Kansas Geological Survey Bulletin 22.* Tulsa, 1936. Pp. 256; illus. 12.

Dr. Moore introduces his work with a statement of a worthy purpose thus:

> It is not the purpose of this report to describe in detail the Pennsylvanian formations of Kansas and adjacent parts of the northern Mid-Continent region, but rather to consider the problems of stratigraphic classification and nomenclature that are needed for description of the system and for treatment of the historical geology of Pennsylvanian time in this region.

The reviewer was, nevertheless, slightly confused as to whether a time or an event basis was fundamental to his delimitation of the boundaries of the Pennsylvanian system of Kansas, or if the boundaries would have been identical on either basis. But multitudinous are the additions to stratigraphic description printed, while seldom is a bit of heterogeneously applied nomenclature redigested in terms of a consistent system, and Dr. Moore certainly deserves credit for notable progress. Moreover, the matter of period boundaries seems to be an unavoidable inconsistency in the face of the present impossibility of long-range correlation. Defining the term "formation" as one of convenience in geologic description (and he distinguishes fundamentally between topographic or lithologic convenience and geologic convenience by applying the principle of the cyclothem to the Kansas deposits which more or less roughly conform to the ideal Pennsylvanian cycles), he proceeds to modify and clarify the use of the formation names, tracing the history of the nomenclature of the major and minor stratigraphic divisions in the Kansas region.

The Pennsylvanian system itself he redefines as including the beds between the Mississippian-Pennsylvanian unconformity (below the Cherokee) and the disconformity above the Brownville limestone and below the Aspinwall limestone, which discontinuity he says represents the Pennsylvanian-Permian break. Dr. Moore regards the present usage in America of the term "Carboniferous" to include the Mississippian, Pennsylvanian, and Permian periods as unfortunate, preferring the European use which does not include the Permian; furthermore, it is his contention that the Mississippian and Pennsylvanian strata should be classified as subdivisions of the Carboniferous system.

JEAN GRACE

ERRATUM

We wish to make a correction in the heading of the review which appears in the *Journal of Geology*, Vol. XLV, No. 6, page 688. It should read as follows:

"Preliminary Report on the Gold Deposits of the Virginia Piedmont," by C. F. Park, Jr., *Virginia Geological Survey Bulletin 44*. Richmond, 1936. Pp. 42; figs. 14; pl. 9.